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Submarine landform assemblages and sedimentary processes in front of Spitsbergen tidewater glaciers

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Abstract

New swath-bathymetric data from the inner parts of the three Svalbard fjords Ymerbukta, Trygghamna, and Magdalenefjorden reveal the landform assemblages deposited in front of tidewater glaciers in west and northwest Spitsbergen. Overridden moraines in Ymerbukta, a tributary of Isfjorden in central west Spitsbergen, record several re-advances of the Esmarkbreen glacier at the head of the fjord after deglaciation, and glacial lineations formed in seafloor sediments are indicative of fast ice advance during one of these events. A terminal moraine and associated debris lobe mark the maximum ice extent during the Holocene, which, implied by the presence of crevasse-squeeze ridges, is likely related to a surge of Esmarkbreen. Several De Geer moraines provide evidence for subsequent slow and step-wise retreat. In the adjacent Trygghamna and in Magdalenefjorden in northwest Spitsbergen the landforms are similar but the absence of overridden moraines and glacial lineations shows that the glaciers probably only re-advanced once during the Holocene and that ice flow was relatively slow. Terminal moraines and associated debris lobes mark the maximum extent of these advances and formed during the Little Ice Age (LIA). In Mag-

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Magdalenefjorden the relatively small size of the debris lobe suggests that the ice margin was at its maximum position for only a short period of time, or that sediment availability was restricted during the LIA advance. Similar to Esmarkbreen the retreat phase of the glaciers in Trygghamna and Magdalenefjorden was also characterised by periods of still-stand or small re-advances, although the comparatively small number of De Geer moraines in all three fjords shows that these landforms probably formed much less frequently than previously thought. Sub-bottom profiler data, four sediment cores and six radiocarbon dates from Magdalenefjorden further provide information about the Holocene sedimentary environments in a northwest Spitsbergen fjord. The main source of sediment is glacial meltwater entering the fjord from the surrounding coastline, which has led to the accumulation of thick sequences of fine-grained mud. Stratified and laminated muds record glacier-proximal conditions, probably related to a LIA re-advance of Waggonwaybreen around 300 cal a BP, where the interplay of a range of glacimarine processes led to the formation of partially rhythmic couplets of one coarser and one finer layer, accumulated at a rate of around 3 cm a^{-1} . Multiple sandy layers intercalated with the glacimarine mud provide evidence for the occurrence of gravitational mass-flow events like turbidity currents. In ice-distal settings, massive to weakly stratified, occasionally bioturbated mud accumulated at a lower rate of $0.04\text{--}0.49 \text{ cm a}^{-1}$. Occasional clasts and diamictic layers show that the depositional environment in Magdalenefjorden is also influenced by sedimentation from icebergs and sea ice, but the ubiquitous glacimarine mud underscores the predominance of meltwater-related sedimentation in the fjord.

Keywords: Fjords, glacial sediments, submarine landforms, west Spitsbergen, Holocene, ice retreat

1. Introduction

High-Arctic glaciers have received increasing attention in the scientific literature as their behaviour is often closely linked to climate change (e.g. Lovell, 2000; Holland et al., 2008; Błaszczyk et al., 2009). In this context, Svalbard has been of particular interest, as around 60% of the archipelago is still covered by glaciers, many of which reach tidewater (e.g. Hagen, 1993; Hagen et al., 2003; Kohler et al., 2007; Mangerud & Landvik, 2007). The ongoing thinning and retreat of these glaciers has resulted in the exposure of well-preserved glacial landform-sediment assemblages in many fjords (e.g. Plassen et al., 2004; Ottesen et al., 2008; Fransner et al., 2017; Streuff et al., 2017), which generally include glacial lineations, terminal moraines marking the maximum extent of ice advance, debris lobes on the distal flanks of the terminal moraines, and recessional moraines or annual push moraines (e.g. Boulton et al., 1996; Plassen et al., 2004; Forwick & Vorren, 2011; Flink et al., 2015; Streuff et al., 2015). Eskers, bedrock highs, crevasse-squeeze ridges, iceberg ploughmarks and pockmarks have further been documented from individual fjords (e.g. Ottesen & Dowdeswell, 2006; Ottesen et al., 2008; Forwick et al., 2009; Dowdeswell & Ottesen, 2016; Flink et al., 2017; Streuff et al., 2017).

The landforms and sediments deposited in front of Svalbard tidewater glaciers serve as an important indicator of past ice dynamics and have led to a better understanding not only of the response of the Svalbard-Barents Sea Ice Sheet (SBIS) to climate change since the Last Glacial Maximum (LGM), but also of individual glacier dynamics throughout the Holocene. The latter are of particular importance for future predictions of the cryosphere as tidewater glaciers can exhibit a wide range of hydrological, thermal, and dynamic regimes (e.g. Blatter & Hutter, 1991; Dowdeswell et al., 1998; Pettersson, 2004). The present study aims to elucidate the individual tidewater glacier dynamics in three Spitsber-

28 gen fjords, two in central west Spitsbergen, and one in northwest Spitsbergen,
29 by using new swath-bathymetric data from the inner fjord basins; it thereby
30 further contributes to the knowledge we have of the glaciation history of the
31 SBIS. Based on the swath-bathymetric data, we identify the landform assem-
32 blages in these fjords and thus complement the work of Forwick et al. (2009),
33 Ottesen & Dowdeswell (2009), and Forwick & Vorren (2012), who documented
34 the landforms in the outer fjord basins. We further use sub-bottom profiler
35 and lithological data, including radiocarbon dates, from Magdalenefjorden to
36 correlate landforms with the sedimentary facies and to reconstruct glacima-
37 rine sedimentary processes and associated accumulation rates. Combined, our
38 data enable us to reconstruct the Holocene depositional processes and prod-
39 ucts in front of selected Spitsbergen tidewater glaciers, to compare and contrast
40 them with those from other glacimarine environments on Svalbard, and thus to
41 evaluate the importance of regional versus local controls on individual glacier
42 dynamics.

43 **2. Study Areas**

44 **2.1. Physiographic Setting**

45 Three fjords were investigated: Ymerbukta, Trygghamna, and Magdalenefjor-
46 den (Figs. 1, 2). They are located on Spitsbergen, the largest island of the
47 Svalbard archipelago. Their physiographic details are summarised in Table 1.
48 Ymerbukta and Trygghamna in central west Spitsbergen are northern tribu-
49 taries of Isfjorden, which, together with its many tributaries, forms Svalbard's
50 largest fjord system (Fig. 1b). Ymerbukta and Trygghamna extend approxi-
51 mately north to south, with the former fjord located east and the latter west
52 of the mountain range Värmlandryggen (Fig. 2a). Both fjords are separated

Table 1: Physiographic details of the fjords from the study area. Local glaciers are listed in the last column with (m) = marine-terminating/tidewater and (t) = terrestrial glaciers. For more details see also Figure 2.

Fjord	Latitude	Longitude	Width [m]	Length [km]	Depth [m]	Glaciers
Ymerbukta (inner basin)	78°16'30"N– 78°18'00"N	13°52'E– 14°05'E	500– 1600	3.5	10–85	Esmarkbreen (m)
Ymerbukta (outer basin)	78°12'00"N– 78°16'30"N	13°52'E– 14°05'E	1300– 2800	6.5	20–160	
Trygghamna (inner basin)	78°14'30"N– 78°16'30"N	13°38'E– 13°52'E	800– 1800	2.4	2–35	Kjerulfbreen (m) Harrietbreen (m)
Trygghamna (outer basin)	78°12'00"N– 78°15'30"N	13°38'E– 13°52'E	1500– 2400	5.5	30–180	Protektorbreen (t) Alkhornbreen (t)
Magdalenefjorden (inner basin)	79°32'30"N– 79°33'30"N	11°08'E– 11°18'E	700	2.5	40–110	Waggonwaybreen (m) Miethebreen (t)
Magdalenefjorden (central basin)	79°33'00"N– 79°34'30"N	10°55'E– 11°05'E	800– 1400	4.5	50–140	Brokebreen (t) Buchanbreen (t)
Magdalenefjorden (outer basin)	79°33'00"N– 79°36'00"N	10°45'E– 10°55'E	900– 4500	7.3	25–95	Adamabreen (t) Skarpeggbreen (t)

into an inner, ice-proximal and an outer, ice-distal basin by a prominent ridge
(Table 1, Fig. 2a).

Magdalenefjorden is located in northwest Spitsbergen (Table 1, Fig. 2b). It
is orientated southeast to northwest and can be subdivided into an innermost
ice-proximal basin, a deep central basin and an outer, ice-distal basin (Table
1, Fig. 2b). Along its southern coast Magdalenefjorden is joined by a small
tributary fjord, Gullybukta, which hosts the glacier Gullybreen at its head (Fig.
2b).

2.2. Previous Work

2.2.1. Glacial Background

During the LGM Svalbard was covered by a large and dynamic ice sheet, which
reached the continental shelf edge and was divided into two components: (1)
fast-flowing ice streams draining the ice sheet through the major fjord systems,
and (2) "inter-ice-stream" areas of slower-moving ice (Elverhøi et al., 1995;
Landvik et al., 1998; Ottesen et al., 2007; Ingólfsson & Landvik, 2013). Is-
fjorden and its tributaries likely served as a major pathway for one of these
large ice streams, with ice reaching its maximum extent at the continental shelf

70 edge between 24 and 20 ka BP (e.g. Mangerud et al., 1992; Svendsen et al.,
 71 1992; Landvik et al., 1998, 2005; Ottesen et al., 2005; Jessen et al., 2010). Ice
 72 retreated rapidly from ~ 14.8 ka BP until ~ 12.3 ka BP, by which time Isfjor-
 73 den’s coastal areas were ice-free (Mangerud et al., 1992; Elverhøi et al., 1995;
 74 Svendsen et al., 1996; Landvik et al., 1998). A re-advance occurred around
 75 12.4 ka BP, after which ice retreated to the heads of the tributary fjords until
 76 ~ 10 ka BP (e.g. Landvik et al., 1995; Ottesen et al., 2007; Forwick & Vorren,
 77 2009). Geomorphological evidence from Ymerbukta and Trygghamna, as well
 78 as lithological evidence from elsewhere in the Isfjorden fjord system, suggest
 79 that the glaciers underwent a significant re-advance during the Younger Dryas
 80 (Mangerud et al., 1992; Svendsen et al., 1996; Forwick & Vorren, 2009, 2011),
 81 which is in contrast to other areas in Spitsbergen where glaciers were smaller
 82 during the Younger Dryas than during the LIA (Mangerud & Landvik, 2007).
 83 In the Isfjorden system, the glaciers experienced asynchronous regrowth after
 84 c. 9 ka BP (e.g. Forwick & Vorren, 2007, 2009; Baeten et al., 2010; Forwick
 85 et al., 2010), and their maximum Holocene extents occurred in response to the
 86 LIA cooling or due to climatically independent surges (e.g. Plassen et al., 2004;
 87 Ottesen & Dowdeswell, 2006; Mangerud & Landvik, 2007; Forwick & Vorren,
 88 2011).

89 In contrast to Isfjorden, Magdalenefjorden was located in an inter-ice-stream
 90 area during the LGM, with ice streams located south and east of the fjord, in
 91 Kongsfjorden and Woodfjorden (Fig. 1b; e.g. Landvik et al., 2005; Ottesen &
 92 Dowdeswell, 2009). Ice flow in Magdalenefjorden was inferred to be relatively
 93 slow due to a more alpine topography in the coastal areas, a thinner ice cover
 94 and generally reduced ice flux (e.g. Landvik et al., 1998; Forman, 1990; Land-
 95 vik et al., 2003, 2005; Ottesen & Dowdeswell, 2009). A grounding zone wedge
 96 close to the continental shelf edge provides evidence that, similar to west Spits-

97 bergen, ice in the northwest reached the shelf edge during the LGM (Ottesen
 98 & Dowdeswell, 2009). The timing of ice retreat from the shelf edge is still
 99 very poorly understood. The only constraints on deglaciation are ^{10}Be expo-
 100 sure dates from the lowlands of the two islands Amsterdamøya and Danskøya
 101 (Fig. 1), which suggest that these areas were ice-free between 18 and 15 ka BP
 102 (Landvik et al., 2005). This may indicate that the mouth of Magdalenefjorden
 103 deglaciated around the same time, suggesting that it was probably ice-free much
 104 earlier than Isfjorden. Indeed, early deglaciation of northwest Svalbard was also
 105 suggested by Forman (1990) and is further supported by Koç et al. (2002) who
 106 placed the disintegration of the northern margin of the Svalbard-Barents Sea
 107 Ice Sheet in northeast Spitsbergen between 13.9 and 13.7 ka BP. From the geo-
 108 morphological evidence ice retreat was inferred to be continuous and relatively
 109 rapid across the outer shelf, and slower but quasi-continuous across the inner
 110 shelf (Ottesen et al., 2007; Ottesen & Dowdeswell, 2009). It is unclear when
 111 Magdalenefjorden was deglaciated and the local tidewater glaciers reached the
 112 head of the fjord, but a minimum deglaciation age of 8.9–8.5 ka BP from the
 113 mouth of the adjacent Smeerenburgfjorden may serve as a rough estimate (Velle,
 114 2012).

115 **2.2.2. Geomorphology**

116 Previous work has been carried out on the landforms in the outer parts of all
 117 three fjords. In outer Ymerbukta Forwick et al. (2009) documented the abun-
 118 dance of single and composite pockmarks, which were described as up to 250 m
 119 wide and 7 m deep. The pockmarks were inferred to have formed from the up-
 120 ward seepage of thermogenic gas, probably predominantly along tectonic faults
 121 occurring in the Isfjorden-Ymerbukta Fault Zone (Dallmann et al., 2002; For-
 122 wick et al., 2009). Indeed, systematic analysis of seafloor seepage features in the
 123 entire Isfjorden fjord system linked the occurrence of pockmarks to the presence

124 of deep bedrock faults and igneous intrusions (Senger et al., 2013; Roy et al.,
 125 2015). Further work by Forwick & Vorren (2012) has shown the presence of
 126 several sediment lobes, as well as blocks and flow structures in both Ymerbukta
 127 and Trygghamna. The majority of the sediment lobes occur in Ymerbukta, are
 128 orientated perpendicular to the main fjord axis, and were described as up to
 129 5-m thick and up to 500-m long and wide deposits. In Trygghamna, where the
 130 presence of sediment lobes is less common, two areas with slide blocks were
 131 identified, where individual blocks are up to 150 m wide, up to 100 m long, and
 132 up to 8 m high (Forwick & Vorren, 2012). The sediment lobes and slide blocks
 133 were inferred to represent debris-flow deposits or slump/slide deposits, formed
 134 from slope failure during the late Holocene with earthquakes and high sediment
 135 accumulation rates as the most likely triggers (Forwick & Vorren, 2012). The
 136 presence of terminal moraines, marking the Holocene maximum extent of the
 137 local glaciers in Ymerbukta and Trygghamna, was also indicated by Forwick &
 138 Vorren (2012). Large sediment lobes on their distal flanks were interpreted as
 139 glacial debris flows formed from high sediment supply to the glacier margin,
 140 causing repeated slope failure during a period of still-stand.

141 Ottesen & Dowdeswell (2009) described the seafloor of Magdalenefjorden as
 142 generally smooth, and related this to sediment infill and draping by silt and
 143 clay settling from suspension in glacial meltwater plumes, with the occurrence
 144 of occasionally protruding bedrock in the form of ridges and knobs. On the steep
 145 side walls of the central fjord basin the same authors described the occurrence
 146 of scars and possibly gullies. In the outer fjord, arcuate sedimentary ridges have
 147 a SW-NE orientation and were identified as moraines formed at the retreating
 148 ice margin during periods of still-stand (Ottesen & Dowdeswell, 2009). A 40
 149 m high ridge across the central fjord was interpreted as the terminal moraine
 150 marking the maximum extent of Waggonwaybreen during the LIA, which occurs

151 in conjunction with several up to 5 m high transverse ridges, interpreted as
 152 push moraines. The latter are thought to form regularly, possibly annually,
 153 during small re-advances of the glacier front during overall ice retreat (Ottesen
 154 & Dowdeswell, 2009). The same processes were inferred for a number of similar
 155 ridges in front of Gullybreen along the southern shore of the fjord (Ottesen &
 156 Dowdeswell, 2009).

157 **2.2.3. Seismostratigraphy**

158 While sub-bottom profiler data and sediment cores are available for Magdalene-
 159 fjorden, and will be investigated here, the seismostratigraphy in Ymerbukta
 160 and Trygghamna is derived from seismic interpretations published by Forwick
 161 & Vorren (2011, 2012), which were based on a chirp profile through Ymerbukta
 162 (F04-017; Fig. 2a), and a boomer profile through Trygghamna (SB-04-023 in
 163 Fig. 2a; for details see Fig. 63.3 in Forwick & Vorren (2012) and Fig. 4b in
 164 Forwick & Vorren (2011)). In addition to the acoustic basement, interpreted
 165 as bedrock, four acoustic units were described from the two fjords: S2, S4a, S5
 166 and S6 (Forwick & Vorren, 2011, 2012). In Isfjorden, at the mouth of the two
 167 fjords, two additional units, S1 and S3, occur.

168 Unit S1 was described as (semi-)transparent with few internal reflections
 169 and a smooth to hummocky upper boundary. It has a draping character, occurs
 170 as basin-fill and on bathymetric slopes, and is present on top of the bedrock in
 171 most of Isfjorden (Forwick & Vorren, 2011). The unit was inferred to represent
 172 sediment deposited subglacially as till or as cavity infill, with a hummocky ap-
 173 pearance ascribed to the presence of glacial lineations (Forwick & Vorren, 2011).
 174 The acoustically massive (semi-)transparent unit S2 was determined to repre-
 175 sent push or thrust moraines accumulating at the glacier front during recession
 176 or minor re-advances during deglaciation, which were inferred to be indicative
 177 of episodic retreat (Forwick & Vorren, 2011). Unit S3 appears as mostly wedge-

178 shaped deposits of a stratified to chaotic acoustic appearance (Forwick & Vorren,
 179 2011). The wedges are usually orientated parallel to the direction of (paleo-)
 180 ice flow and were interpreted as products of (1) glacial outwash, (2) repeated
 181 slope failures caused by high sedimentation rates, isostatic uplift and seismic
 182 activity, or (3) the infill of subglacial cavities (Forwick & Vorren, 2011). The
 183 acoustically stratified deposits of unit S4a are characterised by a draping or on-
 184 lapping geometry and (sub-)parallel internal reflections, and are up to 38 and 40
 185 m thick in Ymerbukta and Trygghamna, respectively (Forwick & Vorren, 2011).
 186 S4a was found to correlate with stratified glacimarine mud with sandy beds and
 187 clasts, which were derived from suspension settling from meltwater, ice raft-
 188 ing, and mass-transport activity in a glacier-proximal environment. The unit
 189 probably dates back to the Younger Dryas and was deposited between 14.1 and
 190 11.2 cal ka BP (Forwick & Vorren, 2009, 2011). Unit S5 represents glacimarine
 191 deposits from the Early Holocene, which are acoustically (semi-)transparent,
 192 conformably overlie S4a, and show occasional onlap geometry. The sediments
 193 were sourced from suspension settling and occasional iceberg rafting during the
 194 Holocene Thermal Optimum between 11.2 and 9 cal ka BP (Forwick & Vorren,
 195 2009, 2011). Unit S6 is the stratigraphically youngest facies in Ymerbukta and
 196 Trygghamna. It is acoustically stratified and interpreted as Holocene glaci-
 197 marine pebbly mud deposited after ~ 9 cal ka BP, when ice rafting increased due
 198 to colder climatic conditions (Forwick & Vorren, 2009, 2011).

199 **3. Methods**

200 For this study, swath-bathymetric data from Magdalenefjorden, Ymerbukta,
 201 and Trygghamna were supplemented with sub-bottom profiler data and four
 202 sediment cores from Magdalenefjorden. The data from Magdalenefjorden were
 203 gathered in October 2009 on the research vessel *R/V Jan Mayen* (now *Helmer*

204 *Hanssen*), using a Kongsberg Simrad multibeam EM300 echo sounder for the
 205 swath-bathymetric, and an EdgeTech 3300-HM sub-bottom profiler for the chirp
 206 data. Swath-bathymetric data from Ymerbukta and Trygghamna were gath-
 207 ered in July 2000 by the Norwegian Hydrographic Survey with a Kongsberg
 208 Simrad multibeam EM1002 installed on the research vessel *Sjømaleren* (now
 209 *IXPLORER*). Both multibeam echo sounders were calibrated using frequent
 210 conductivity-temperature-depth (CTD) measurements from the water column.
 211 Bathymetric data were gridded to a cell size of 5x5 m in DMagic, and visualised
 212 and interpreted in the Fledermaus v7 3D Software. The chirp sub-bottom pro-
 213 filer was operating at a pulse mode of 2–16 kHz and 3 ms during acquisition
 214 and the resulting data were processed using the EdgeTech Software and vi-
 215 sualised and interpreted in the Kingdom Suite software. Generally assuming
 216 water-saturated sediments, an average p-wave velocity of 1500 m s^{-1} was used
 217 for the correlation between seismo- and lithostratigraphy, i.e. conversion from
 218 ms to m.

219 A total of four gravity cores, JM09H-GC01, GC03, GC06, and GC09, were
 220 taken from Magdalenefjorden, using a 1900 kg heavy gravity corer with a 6 m
 221 long steel barrel and an inner diameter of ~ 110 mm. All cores were cut into sec-
 222 tions up to 1.3 m in length, split into working and archive halves, and stored at
 223 $+4^\circ\text{C}$. It is important to note that GC01 and GC03 were split in 2009, whereas
 224 GC06 and GC09 were split in 2016. Thus, certain parameters (e.g. water con-
 225 tent) may not be directly comparable between the four cores, although variations
 226 within one core are informative. In order to identify the different lithofacies in
 227 the fjord, core logs were generated from the visual description of the sediment
 228 surface of the working halves, which were supplemented using the information
 229 on internal sedimentary structures provided by x-radiographs. The latter were
 230 acquired with a GEOTEK Thermo Kevex PSX10-65W-Varian2520DX, running

231 with a voltage of ~ 95 kV and a current of around 150 nA. All cores were run
 232 through a GEOTEK multi-sensor core logger at Durham University to deter-
 233 mine the wet-bulk density and magnetic susceptibility in 1 cm intervals, the
 234 latter of which was measured with a Bartington point-sensor mounted on the
 235 system. Magnetic susceptibility values were obtained in SI units ($\chi \times 10^{-5}$) but
 236 are displayed unitless throughout this article. Bivalve shells located from the
 237 x-rays of GC06 and benthic foraminifera from selected sediment depths in GC01
 238 were sent to Beta Analytic for Accelerator Mass Spectrometry (AMS) dating.
 239 The obtained conventional ^{14}C ages were calibrated with the MARINE13 cal-
 240 ibration (cf. Reimer et al., 2013), using a marine reservoir effect of 400 years
 241 and a local ΔR of 61 ± 70 years (Mangerud, 1972). For the water content 1
 242 cm thick sediment slabs were taken at selected (usually lithofacies-dependent)
 243 core depths, weighed, dried at 60°C , and weighed again. The same samples
 244 were subsequently soaked overnight in a solution of distilled water and sodium-
 245 hexametaphosphate to encourage the disintegration of flocculated clay particles.
 246 The sediments were then sieved through mesh sizes of 500, 250, 125 and $63 \mu\text{m}$
 247 (stacked on top of each other) to determine the grain size distribution within
 248 the cores.

249 **4. Results**

250 **4.1. Geomorphology**

251 Based on the swath-bathymetric data from the inner parts of Ymerbukta, Tryg-
 252 ghamna, and Magdalenefjorden we identify a total of five different landform
 253 types. The distribution of all observed landforms is shown in Fig. 3, which incor-
 254 porates those landforms previously described by Forwick et al. (2009), Ottesen
 255 & Dowdeswell (2009), and Forwick & Vorren (2012).

256 **Groove-ridge features – glacial lineations**

257 Elongate, (sub-)parallel ridges and grooves are present in inner Ymerbukta (Fig.
258 4a, b). These features are 25 m high, 150–800 m long, up to 20 m wide and
259 spaced at irregular distances between 50 and several hundred meters (Fig. 4a,
260 b). The crests are straight to slightly curved, and round and symmetrical in
261 cross-section. These ridges are orientated parallel to the main fjord axis, i.e. in
262 the direction of ice flow (Fig. 3b). The majority of these features are overprinted
263 by transverse ridges (see section 4.1 below), indicating that the elongate ridges
264 were deposited first. Similar, albeit larger ridges (up to 2 km long, ~200 m
265 wide) appear in Isfjorden, where their orientation is northeast-southwest, i.e.
266 almost normal to the ridges in Ymerbukta (Fig. 3b).

267 Similar sets of ridges and grooves have previously been described from other
268 Spitsbergen fjords, where they were interpreted as glacial lineations (e.g. Otte-
269 sen & Dowdeswell, 2006; Ottesen et al., 2008; Streuff et al., 2015). Indeed,
270 Ottesen et al. (2007) documented the presence of (mega-scale) glacial lineations
271 in Isfjorden. Glacial lineations are formed beneath a glacier or ice stream, where
272 the soft subglacial sediments are deformed into ridges and grooves by processes
273 of erosion and re-deposition (e.g. Tulaczyk et al., 2001; Ó Cofaigh et al., 2005).
274 They are commonly associated with fast ice flow (King et al., 2009).

275 **Transverse ridges – recessional moraines**

276 Several large ridges in the innermost fjord basins are orientated transverse to the
277 main fjord axis (Figs. 3, 4) and extend across the width of the fjords. The ridges
278 are up to 20 m high, 800–2000 m long and 100–300 m wide. Their sometimes
279 multiple crests are sinuous in planform, round, and mostly symmetrical in cross-
280 section (Fig. 4d, f). In Ymerbukta some of these ridges are overprinted by the
281 glacial lineations. The most distal ridges in Ymerbukta and Trygghamna are

282 associated with large sediment lobes (see section 2.2.2), which extend down their
 283 distal flanks (Fig. 3b). In Isfjorden a large feature occurs, which is similar to the
 284 transverse ridges in the fjords, but is ~ 85 m high, ~ 4 km wide and more than
 285 3.8 km long (Fig. 4h-j). It is characterised by an asymmetric cross-profile with
 286 a steeper distal and a flatter proximal side (Fig. 4). In plan-view, it appears
 287 more as a plateau with a sharp edge rather than a ridge with a crest (Figs. 3a,
 288 b, 4).

289 The large transverse ridges are interpreted to be recessional moraines de-
 290 posited during overall glacier retreat. This is in accordance with e.g. Ottesen &
 291 Dowdeswell (2006), Baeten et al. (2010) and Streuff et al. (2015), who described
 292 similar ridges from other Spitsbergen fjords. In Ymerbukta, where these ridges
 293 are partially streamlined or overprinted by glacial lineations, they probably rep-
 294 resent recessional moraines from an earlier advance of Esmarkbreen, that were
 295 overprinted by a later re-advance. The transverse ridges at the outermost edge
 296 of the inner basins in Ymerbukta, Trygghamna, and Magdalenefjorden have
 297 previously been interpreted as terminal moraines, deposited when the glaciers
 298 reached their maximum Holocene extents, either in response to the LIA cool-
 299 ing (Ottesen & Dowdeswell, 2009) or as a result of a surge (Forwick & Vorren,
 300 2012).

301 We suggest that the ridge-like feature in Isfjorden formed as a sedimentary
 302 wedge at or close to the grounding line of an ice stream during a period of still-
 303 stand during overall ice retreat. Although formation as a grounding-zone wedge
 304 could be feasible in this context, a seismic profile across the feature (SS97-163;
 305 Fig. 4j) does not show the characteristic reflection pattern (cf. Forwick & Vor-
 306 ren, 2011; Batchelor & Dowdeswell, 2015). We thus support the interpretation
 307 of Forwick & Vorren (2011), who, from the seismic profile, inferred the forma-
 308 tion of wedge-shaped ice-marginal deposits either from repeated slope failure,

309 as glacial outwash or as sedimentary infill of subglacial cavities.

310 **Small transverse ridges – De Geer moraines**

311 Abundant small transverse ridges occur in the inner fjord basins, which extend
312 across the fjords and are between 100 and 2000 m long, around 20 m wide,
313 and ~ 3 m high. These ridges overprint the glacial lineations in Ymerbukta,
314 indicating that they are the youngest features. The majority have sharply de-
315 fined, continuous crests, are largely symmetrical in cross-section, and are spaced
316 at intervals between 10 and 50 m (Fig. 4a, c). In places these ridges exhibit
317 branching and may cross-cut each other (Fig. 3).

318 The small ridges are similar to annual push moraines described from other
319 Svalbard fjords (e.g. Ottesen et al., 2008; Flink et al., 2015), which are formed
320 each year during winter, when the presence of shore-fast sea ice causes the gen-
321 erally retreating glaciers to stagnate or even re-advance (Boulton, 1986; Ottesen
322 & Dowdeswell, 2006). The overall regular spacing supports this interpretation.
323 The fact that these ridges exhibit branching and cross-cut each other in places
324 could be indicative of other processes involved, e.g. the squeezing of soft sub-
325 glacial sediments into basal glacier crevasses (see Evans & Orton, 2014; Streuff
326 et al., 2015). This is supported by the presence of terrestrial crevasse-squeeze
327 ridges in the forefield of Esmarkbreen (Farnsworth et al., 2016). Given the likely
328 mixed origin of the small ridges both as end moraines and as crevasse-squeeze
329 ridges, we interpret these landforms as De Geer moraines (cf. Lundqvist, 1981;
330 Streuff et al., 2015).

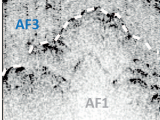
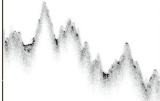
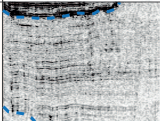
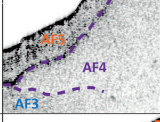
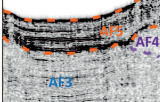
331 4.2. Seismostratigraphy

332 Description

333 The chirp data available from Magdalenfjorden reveal a total of five acoustic
334 facies, AF1–AF5 (Fig. 5). Their characteristics are summarised in Table 2.

335 AF1 is the stratigraphically lowermost facies in Magdalenefjorden and is de-
336 fined by very weak and chaotic internal reflections that disappear with depth
337 and a discontinuous, often hummocky upper boundary (Table 2, Fig. 5). AF2
338 directly overlies AF1 in most of the fjord and has a similar acoustic signature
339 to AF1 (Table 2, Fig. 5c). The top boundary of AF2 is usually strong and
340 opaque and is sometimes accompanied by strong diffraction hyperbolae (Fig.
341 5c). AF3 conformably overlies AF2, or AF1, where AF2 is absent. Its acoustic

Table 2: Acoustic facies in Magdalenefjorden. An average p-wave velocity of 1500 m s^{-1} was used to convert thickness from ms to m.

ID	Acoustic signature	Bounding reflectors	Geometry/Thickness	Distribution	Interpretation
AF1	 AF1	Lower: not detected Upper: hummocky, variably strong, discontinuous	Thickness: >10 ms (≈7.5 m)	As acoustic basement in entire fjord	Bedrock
AF2	 AF2	Lower: discontinuous, hummocky, variably strong Upper: strong, mostly continuous, hummocky in inner fjord basin	Thickness: 1-20 ms (≈1.75-15 m)	Overlying AF1 in most of the fjord, also occurs as S2 in both Ymerbukta and Trygghamna	Glacial till
AF3	 AF3	Lower: variably strong, (dis-)continuous Upper: strong, opaque, mostly continuous	Conformable geometry Thickness: <24 ms (≈18 m)	Common in basins, overlies AF1 or AF2; also occurs in Ymerbukta and Trygghamna as S4a	Gacimarine mud from suspension settling; occasionally coincide with mass flows
AF4	 AF4	Lower: weak, mostly continuous Upper: continuous, variably strong	Occurs as lenses or wedge-shaped deposits that laterally pinch out Thickness: 2-22 ms (≈1.5-16.5 m)	Central fjord, common at the foot of slopes, often appears intercalated into AF3	Turbidity currents or alternative mass-flow deposits (MTDs)
AF5	 AF5	Lower: continuous, variably strong Upper: seabed	Conformable geometry, overlies AF4 Thickness: <6 ms (≈4.5 m)	Inner fjord, also occurs in Ymerbukta and Trygghamna as S6	Late Holocene glaci-marine or hemipelagic mud; occasionally coincides with MTDs

342 appearance varies from internally massive and semi-transparent to acoustically
 343 stratified (Table 2, Fig. 5c). It has a strong and opaque, relatively continu-
 344 ous upper boundary and is especially common in bathymetric basins (Fig. 5b,
 345 c). AF4 is acoustically (semi-)transparent with weak internal reflections of a
 346 massive nature that disappear in places (Table 2, Fig. 5c). The facies occurs
 347 as lenses or wedge-shaped deposits, often intercalated into AF3 (Fig. 5) and
 348 commonly appears at the foot of slopes where it pinches out with increasing
 349 distance from the slope (Fig. 5c, d). AF5 is the stratigraphically uppermost
 350 facies in Magdalenefjorden, and thus the inferred youngest. It is bounded by the
 351 seabed on top and appears acoustically stratified (Fig. 5c). AF5 is present in
 352 the inner fjord, i.e. ice-proximal areas, and conformably overlies AF4 (Fig. 5).

353 **Interpretation**

354 AF1 is interpreted as the acoustic basement in Magdalenefjorden, the acoustic
 355 appearance of which suggests that AF1 represents either glacial till or bedrock
 356 (cf. Forwick et al., 2010; Forwick & Vorren, 2012; Kempf et al., 2013; Roy
 357 et al., 2014). Its stratigraphic position, the hummocky upper boundary, and
 358 the outcropping bedrock visible on the swath-bathymetric data (Fig. 5) all
 359 indicate that the acoustic basement in Magdalenefjorden comprises bedrock.

360 AF2 is acoustically similar to AF1 and to acoustic facies from other Spits-
 361 bergen fjords, including Unit S2 in Ymerbukta and Trygghamna (cf. Forwick &
 362 Vorren, 2011, 2012; Kempf et al., 2013) and is thus interpreted as glacial till.
 363 This would imply a diamictic composition for the sediments of AF2, which is
 364 supported by the massive acoustic appearance and weakening internal reflections
 365 with depth (cf. Stewart & Stoker, 1990; Forwick & Vorren, 2012). In innermost
 366 Magdalenefjorden AF2 crops out at the seafloor and its very hummocky appear-
 367 ance here correlates with the occurrence of the recessional moraines (Fig. 5),
 368 also supporting an interpretation as glacial till.

369 The (semi-)transparent, partially stratified acoustic signature of AF3 sup-
 370 ports an interpretation as glacimarine sediments derived from suspension set-
 371 tling from meltwater plumes (e.g. Plassen et al., 2004; Kempf et al., 2013; Streuff
 372 et al., 2015). Where the facies appears stratified, regular changes in lithology or
 373 density are implied, which could indicate a glacier-proximal depositional envi-
 374 ronment (cf. Syvitski, 1989; Forwick & Vorren, 2012), possibly with the occur-
 375 rence of turbidity currents or other gravitational mass flows (e.g. Sexton et al.,
 376 1992; Forwick & Vorren, 2012; Streuff et al., 2017). In places the latter is also
 377 indicated by the localised appearance of AF4 (see below). AF3 is similar to S4a
 378 in Ymerbukta and Trygghamna (cf. Forwick & Vorren, 2011, 2012).

379 AF4 is interpreted as mass-flow deposits. This is supported by the lenticular
 380 appearance, the internally massive semi-transparent acoustic signature, and its
 381 similarity with such deposits in other areas around Svalbard (e.g. Plassen et al.,
 382 2004; Forwick & Vorren, 2007; Hogan et al., 2010; Forwick & Vorren, 2012;
 383 Kempf et al., 2013; Streuff et al., 2015).

384 AF5 is interpreted as late Holocene glacimarine or hemipelagic sediments
 385 delivered into the fjord by meltwater streams (cf. S6 in Ymerbukta and Tryg-
 386 ghamna, AF6 in Lomfjorden; Forwick & Vorren, 2011; Streuff et al., 2017).
 387 Similar to S5 and S6, the internal reflections could indicate periods of increased
 388 iceberg rafting (cf. Forwick & Vorren, 2009, 2011). However, our lithological
 389 data suggest that the stratification is probably the result of repeated changes
 390 in grain size and possibly mass flow activity (see section 4.3 below).













391 **4.3. Lithostratigraphy**

392 **4.3.1. Lithofacies**

393 **Description**

394 Based on the sedimentary evidence, we define a total of four lithofacies, LF1–

Table 3: Lithofacies in Magdalenefjorden with examples of the facies in core logs and on x-radiographs. For physical properties MS = Magnetic Susceptibility, WC = Water Content, and ρ = wet-bulk density as measured with the MSCL.

Litho-facies	Log Clay Silt Sand	X-ray	Material	Physical parameters	Distribution	Interpretation
LF1			massive, greyish brown, no internal structures, very few clasts, occasionally bioturbated	2.5Y 5/2 - 4/2 >90% clay + silt MS \approx 20 WC = 25-30% $\rho \approx 1.8-2.1$ g/cm ³	Upper 25 cm of GC01 and GC03 (Fig. 6)	Distal glaci-marine mud settling from suspension in glacial meltwater plumes and the water column
LF2a			weakly stratified, partially bioturbated; occasional clasts and load structures	2.5Y 5/2 - 4/2 >98% clay + silt	GC01, GC03 (Fig. 6)	Distal glaci-marine mud settling from suspension in meltwater
LF2b			stratified, couplets of sharp-bounded coarse and fine layers, occasionally with LF4	MS \approx 10-20 WC = 20-35%		Proximal glaci-marine mud, deposition from suspension probably seasonally controlled
LF2c			(diffusely) stratified mud with thin diamict layers, gradual contacts	$\rho \approx 1.9-2.25$ g/cm ³		Glaci-marine mud with IRD, deriving from suspension and icebergs/sea ice
LF3			diffusely laminated, dark grey at surface, black in sub-surface, strong smell of H ₂ S, commonly shell fragments and bivalves	5Y 4/1 5Y 2.5/1 40-80% clay + silt MS \approx 5-10 WC = 20-35% $\rho \approx > 2$ g/cm ³	GC06 (Fig. 6)	Distal glaci-marine mud settling from suspension in glacial meltwater plumes from different sources. High organic content likely cause of seasonal control on biological productivity
LF4			massive sandy mud, occurs as thin beds or thick strata, strong smell of H ₂ S, black sub-surface, few shell fragments, sharp contacts	2.5Y 4/2 - 5/2 5Y 2.5/1 60-80% clay + silt MS \approx 15-30 WC = <25% $\rho \approx 2-2.5$ g/cm ³	GC03, GC06, GC09 (Fig. 6)	Glacier outwash or sediment remobilised during gravitational mass flows

LF4 in Magdalenefjorden (Fig. 6). Their characteristics are summarised in Table 3.

LF1 contains massive mud with > 90% clay and silt and only occurs in the top 25 cm of GC01 and GC03 (Table 3). It shows no internal structures and very few clasts (Fig. 6). The water content (Table 3) is a minimum value, as both cores were split in 2009, and it is likely that the sediments dried out to

401 some extent since then.
 402 *LF2* contains finely laminated to stratified mud with occasional clasts and oc-
 403 curs in GC01 and GC03. Like in *LF1*, the mud is generally very fine (Table
 404 3, Fig. 6). Black mottles appear throughout the facies. We define laminae as
 405 thinner than 1 cm (but usually up to only 2 mm in Magdalenefjorden), and
 406 strata, which are between 1 and 5 cm thick. The greyscale changes in the x-
 407 radiographs suggest that the stratified nature of the sediments is likely caused
 408 by differences in grain size or density (Fig. 6c, f). The latter is supported by the
 409 strong oscillations in bulk density (Fig. 6). The water content, as for *LF1*, is a
 410 minimum value. We distinguish three subfacies of *LF2*: *LF2a*, *LF2b*, and *LF2c*.
 411 *LF2a* contains weakly or diffusely stratified, partially bioturbated mud where no
 412 significant changes in grain size occur between individual layers (Fig. 6f). Occa-
 413 sional clasts within the facies may be accompanied by load structures. In *LF2b*
 414 the stratification is more pronounced and defined by couplets of sharp-bounded
 415 coarser and finer layers of variable thickness (Fig. 6f). Bioturbation traces are
 416 absent. Layers may be inclined with undulating contacts and contain the fine
 417 massive sand of *LF4* (see below) in places. *LF2c* contains (diffusely) stratified
 418 mud, which appears intercalated with thin layers of a matrix-supported diamict
 419 (Fig. 6c, f). The boundaries between mud and diamict are gradual and lack
 420 load structures (Fig. 6f).
 421 *LF3* only occurs in GC06 and contains diffusely laminated mud with a relatively
 422 high silt and sand content (Table 3, Fig. 6). The sediments are dark grey at
 423 the surface, and black in the subsurface, and emit a strong smell of H_2S . Shell
 424 fragments and entirely preserved bivalves are common and may occur in con-
 425 centrated layers or scattered throughout (Fig. 6). The slightly coarser nature of
 426 *LF3* is reflected in the grain size distribution, which commonly shows 40–80%
 427 of the sediments to be finer than $63\ \mu m$ (Table 3, Fig. 6). Although the water

content of 20–35% is similar to the fine-grained mud of LF2a, it is important to note that GC06 was one of the two cores split in 2016, its water content is thus regarded as more reliable, and the water content of LF2a may consequentially be higher than measured.

LF4 occurs as thin beds or as thick strata and has a massive, unsorted appearance. It contains fine sand with small, but variable amounts of mud and sparsely scattered clasts and shell fragments (Table 3). During core splitting a strong H₂S odour occurred and the sediments’ subsurface is characterised by a black colour. Boundaries to other lithofacies are often sharp (Fig. 6). GC09 is only composed of LF4 (Fig. 6b).

Interpretation

LF1 is similar to massive muds observed in other glacimarine settings and is interpreted as mud settling from suspension in meltwater plumes or the water column. The massive internal structure and the presence of occasional bioturbation structures indicate ice-distal conditions during the accumulation of LF1 (cf. e.g. Elverhøi et al., 1983; Forwick & Vorren, 2009; Baeten et al., 2010).

The stratified muds of LF2 are similar to sediments documented from other fjords in Spitsbergen and are accordingly interpreted as glacimarine mud deposited in a relatively calm, meltwater-dominated depositional environment (cf. e.g. Elverhøi et al., 1983; Plassen et al., 2004; Forwick et al., 2010; Streuff et al., 2015). Laminated or stratified glacimarine muds commonly occur in Arctic fjords, where they are usually associated with an interplay of depositional processes or sedimentation from multiple sources (cf. e.g. Elverhøi et al., 1980; Ó Cofaigh & Dowdeswell, 2001; Forwick & Vorren, 2009; Streuff, 2013). Stratification and lamination of LF2 and its subfacies in Magdalenefjorden could thus be a consequence of (1) multiple glaciers delivering sediments to the source, (2) tidally controlled release of the suspension load carried in the meltwater plumes,

(3) diurnal or seasonal variation in meltwater flux and thus depositional energy, and (4) an interplay between suspension settling from meltwater plumes and repeated gravity flows (e.g. Elverhøi et al., 1983; Mackiewicz et al., 1984; Cowan & Powell, 1990; Ó Cofaigh & Dowdeswell, 2001; Gilbert et al., 2002). This is further discussed in section 5.2 below.

LF3 is similar to LF2a with the main differences being darker sediments and coarser grains in LF3. Above, we suggested a number of processes that could lead to the stratified appearance of glacial-marine mud for the formation of LF2, and infer that similar processes deposited LF3 (see also section 5.2 below). Black mottles and the smell of H_2S are suggested to reflect monosulphid layers, formed from seasonal variations in biological productivity (e.g. Elverhøi et al., 1980, 1983).

The unsorted and massive appearance of LF4 as well as the coarser grain sizes are interpreted as the result of sediment accumulation in a high-energy depositional environment, i.e. from gravity flows. Where LF4 occurs as thin beds or lenses, these flows were probably of relatively small magnitude and could represent turbidite deposits in a glacier-proximal environment (cf. e.g. Gilbert, 1982; Gilbert et al., 1993). Larger packages of LF4, such as at the top of GC06, may accordingly result from more extensive mass-transport deposits, which commonly occur in Svalbard fjords, especially along the fjord walls (e.g. Forwick & Vorren, 2007, 2012; Streuff et al., 2017, this study).

4.3.2. Radiocarbon dates and sediment accumulation rates

Six radiocarbon dates were obtained from foraminifera and molluscs from GC01 and GC06 in Magdalenefjorden. The dates are displayed in Table 4 and Figure 7.

The results from the AMS dating show that the sediments in GC01 reflect recent deposition, with a basal date of ~ 505 cal a BP, an age of ~ 319 cal a BP

Table 4: Radiocarbon dates from this study, calibrated using a ΔR of 61 ± 70 (Mangerud, 1972). Note that the reported age for Beta-434937 indicates an age of post 1950 AD and is therefore given in percent modern carbon (pMC) of the modern reference standard.

Core ID	Depth [cm]	Lab Code	Sample	Reported age [^{14}C a BP]	Median [cal a BP]	2σ [cal a BP]
GC01	25	Beta-447123	Foraminifera	660 ± 30	228	0–20; 49–410
GC01	254	Beta-447124	Foraminifera	730 ± 30	319	131–472
GC01	359–363	Beta-448203	Foraminifera	950 ± 30	505	360–640
GC06	3.5	Beta-434937	Bivalve	$106.2 \text{ pMC} \pm 0.3$	NA	~ 0 –60
GC06	12	Beta-434938	Single valve	560 ± 30	132	0–262
GC06	35	Beta-434939	Fragmented bivalve	550 ± 30	124	0–257

for a sediment depth of 254 cm, and an age of ~ 228 cal a BP at a sediment depth of 25 cm (Table 4, Fig. 7). As all dates stem from around 6 mg of generally well-preserved benthic foraminifera we consider the foraminifera in GC01 to reflect in-situ conditions and infer that the obtained ages are reliable. Assuming a linear rate of sediment accumulation, the dates from GC01 suggest that LF2a in the lower parts of the core accumulated at a rate of $\sim 0.49 \text{ cm a}^{-1}$, whereas LF2b, between 254 and ~ 50 cm sediment depth, accumulated at a rate around 3.3 cm a^{-1} . LF1 in the topmost 25 cm of the core was deposited at a sediment accumulation rate (SAR) of 0.04 cm a^{-1} (Fig. 7). The relatively high SAR in the central parts of the core suggests that an interpretation of LF2b as ice-proximal sediments is reasonable (cf. e.g. Elverhøi et al., 1980, 1983). Similarly, the lower SAR of the topmost 25 cm supports a distal glacimarine environment. The sediments in GC06 are younger than those in GC01, as the AMS dating shows ages of ~ 124 cal a BP for a sediment depth of 35 cm, ~ 132 cal a BP for a sediment depth of 12 cm, and an age of <60 years for the sediments around 3.5 cm core depth. The age reversal between the two lowermost dates in GC06 may indicate that the single valve at 12 cm was reworked, but considering the very young ages of the sediments at both depths, the large error range may easily account for the older age at 12 cm. Furthermore, the weakly stratified nature

501 of LF3 and the good preservation of the shells sampled suggest that sediment
 502 reworking is unlikely in this part of the core. The two ages at 12 and 35 cm
 503 could thus suggest that the shells were deposited simultaneously and that LF3
 504 in GC06 accumulated quasi-instantaneously. However, as there is some evidence
 505 of bioturbation in the core, it is also possible that the age reversal at 12 cm was
 506 caused by post-depositional upward migration of an older shell into younger
 507 sediments. If this were the case, LF3 was deposited at a SAR of 0.25–0.49 cm
 508 a^{-1} in GC06, indicating that this lithofacies was probably deposited in a similar
 509 environment to LF2a.

510 **5. Discussion**

511 **5.1. Landform assemblages in Ymerbukta, Trygghamna and** 512 **Magdalenefjorden**

513 The landforms described from Ymerbukta, Trygghamna, and Magdalenefjor-
 514 den include (1) bedrock highs, (2) overridden recessional moraines, (3) glacial
 515 lineations, (4) terminal moraines, (5) deposits from gravitational mass flows,
 516 (6) unmodified recessional moraines, (7) De Geer moraines, and (8) pockmarks.
 517 While there is some variability in the landforms observed in the outer fjord
 518 basins, the landform assemblages in the ice-proximal basins adjacent to the
 519 present glacier fronts are remarkably similar. In Ymerbukta and Trygghamna
 520 the more distal fjord basins are characterised by a relatively smooth seafloor
 521 with pockmarks and debris flow deposits as the only landforms. In Magdalene-
 522 fjorden, pockmarks and mass flow deposits are absent and bedrock highs and
 523 partially buried recessional moraines are the only features observed. The smooth
 524 seafloor in all three fjords suggests some degree of sediment draping, possibly
 525 masking underlying landforms (cf. Ottesen & Dowdeswell, 2009), which is con-

526 firmed by the thick sedimentary sequences observed on the sub-bottom profiler
527 data (Forwick & Vorren, 2011, 2012, this study). The presence or absence of
528 debris flows could be related to local differences in sediment availability, ac-
529 cumulation rates, exposure to earthquakes, and/or the presence or absence of
530 gaseous fluids beneath the seafloor (cf. e.g. Hovland & Judd, 1988; Forwick
531 et al., 2009; Forwick & Vorren, 2012).

532 Based on the presence of the landforms in the inner fjord basins and their
533 relationship with each other, we define three landform assemblages, one for
534 each fjord. All assemblages are named after their inferred source glacier. The
535 Esmarkbreen Assemblage occurs in Ymerbukta and comprises overridden re-
536 cessional moraines, glacial lineations, a terminal moraine and associated debris
537 lobes on its distal flank, and a number of De Geer moraines. The Kjerulfbreen-
538 Harrietbreen Assemblage in Trygghamna is made up of a terminal moraine and
539 an associated debris lobe, three recessional moraines, and a sequence of De Geer
540 moraines. The three recessional moraines and the De Geer moraines in inner
541 Magdalenefjorden form the Waggonwaybreen Assemblage. Their formation is
542 discussed in section 5.3 below.

543 **5.2. Sedimentary environments**

544 From the lithological data in Magdalenefjorden, we identify three main sedimen-
545 tary processes in the fjord: (1) suspension settling of fine-grained glacimarine
546 mud from glacial meltwater, (2) delivery of clasts to the fjords as ice-rafted de-
547bris melting out from icebergs and sea ice, and (3) abundant gravitational mass
548 flows reworking the accumulated sediments. The first process, i.e. the rain-out
549 of suspension load from meltwater, is the most dominant process in Magdalene-
550 fjorden, where the seafloor is covered by a relatively thick sequence of glaci-
551 marine muds. These muds are massive at the top of GC01 and GC03 (LF1), where

552 they overlie the (diffusely) stratified mud of LF2. While the depositional process
 553 is the same for both lithofacies, the different internal structure indicates slight
 554 changes in the depositional environment. Massive mud as recorded in LF1 could
 555 reflect more distal conditions, in which there is insufficient energy to transport
 556 coarser grains to the core sites and high-energy gravitational processes are rare.
 557 Ice-distal conditions and thus lower energy during the accumulation of LF1 are
 558 further supported by the low SAR of 0.04 cm a^{-1} and the scattered biotur-
 559 bation traces (cf. e.g. Elverhøi et al., 1980; Lønne & Mangerud, 1991; Sexton
 560 et al., 1992; Baeten et al., 2010). Similar conditions are inferred for the diffusely
 561 stratified sediments of LF2a. In contrast, the more pronounced lamination or
 562 stratification characteristic of LF2b is considered to be consistent with a more
 563 glacier-proximal setting (e.g. Elverhøi et al., 1983; Ó Cofaigh & Dowdeswell,
 564 2001; Forwick & Vorren, 2009), where individual strata were formed due to
 565 regular variations in depositional energy. Thinner, more rhythmic laminae, for
 566 instance, probably derive from semi-regular changes in meltwater supply or
 567 suspension release from the meltwater plumes, which could be tidally or sea-
 568 sonally controlled (e.g. Elverhøi et al., 1980; Mackiewicz et al., 1984; Cowan
 569 et al., 1997; Ó Cofaigh & Dowdeswell, 2001). Given the sedimentation rate of
 570 3.7 cm a^{-1} in GC01 we can infer that two such couplets occur per year (in
 571 4 cm of sediment), providing evidence that some of these laminations are in-
 572 deed seasonally controlled. The abundance of black mottles in LF2 supports
 573 this, as it is possibly related to seasonal variations in biological productivity,
 574 where black layers reflect diatom blooms in spring (Elverhøi et al., 1980, 1983).
 575 Where the lamination or stratification is more irregular, and sandy beds oc-
 576 cur, the deposition of glacial marine mud from meltwater is probably influenced
 577 by punctuated single events, such as turbidity currents. This is supported by
 578 the sharp-based, sometimes undulating boundaries of the sand layers and their

579 downward inclination (see Fig. 6f), which are common features of turbidite
 580 deposits from glacier-proximal environments (e.g. Kuenen, 1948; Gilbert, 1982;
 581 Powell & Molnia, 1989; Gilbert et al., 1993). The relatively high SAR around
 582 3 cm a^{-1} and the absence of bioturbation traces within this facies further sup-
 583 port a glacier-proximal origin for LF2a. The clasts scattered throughout the
 584 glacialine muds in Magdalenefjorden likely represent IRD and therefore imply
 585 that sedimentation from icebergs and sea ice is another process influencing the
 586 depositional environment in the fjord. However, the relatively low number of
 587 clasts emphasises the importance of sedimentation from meltwater, which partly
 588 masks the input from IRD. Individual clasts in lithofacies LF2a are probably
 589 dropstones, as indicated by the occasional presence of load structures. The rare
 590 diamictic and clast-rich layers embedded into the weakly stratified mud of LF2c
 591 on the other hand probably originate from concentrated melting events, either
 592 related to iceberg turnover or to the meltout of IRD from shore-fast sea ice and
 593 entrapped icebergs during summer, while the surrounding mud was deposited
 594 during winter when IRD supply was suppressed (cf. Syvitski et al., 1996; Jen-
 595 nings & Weiner, 1996; Dowdeswell et al., 2000). Considering the nearly stratified
 596 appearance of the diamict layers, the gradual, quasi-horizontal contacts (see Fig.
 597 6f), and the absence of load clasts, seasonally-controlled sedimentation appears
 598 to be more likely than iceberg dumping.

599 Similar to LF2a, LF3 is interpreted as relatively ice-distal glacialine mud.
 600 The colour difference between the two facies could indicate a different melt-
 601 water source, i.e. variable sediment provenance (cf. Forwick & Vorren, 2009;
 602 Forwick et al., 2010), an assumption that seems reasonable when considering
 603 the location of the core sites. GC01 and GC03, which predominantly contain
 604 LF1, are located in the inner and central fjord and are thus likely influenced by
 605 meltwater supplied by the four glaciers at the head of the fjord, Buchanbreen,

606 Waggonwaybreen, Miethebreen, and Brokebreen. The core site of GC06 in the
607 outer fjord, on the other hand, may additionally be influenced by sediment sup-
608 ply from Gullybreen, Adambreen, and Skarpeggbreen (see Fig. 2). The strong
609 smell of H_2S and the generally darker sediment colour is probably related to
610 a relatively high organic content, which, again, could contribute to the diffuse
611 stratification in the form of annual monosulphid layers formed by spring algal
612 bloom (Elverhøi et al., 1980, 1983).

613 The massive sand intermixed with glacial marine mud in LF4 is interpreted as
614 a product of glacier outwash, especially in the inner fjord, or of larger mass-
615 transport events. The latter is more likely, as the frequent occurrence of such
616 events in all fjords is also attested to by the sedimentary lobes visible on the
617 bathymetric data and by the appearance of the lenticular bodies of AF4 and
618 S4a, which were at least partially interpreted as gravity-flow deposits (Forwick
619 & Vorren, 2011, this study). The fact that GC09 and GC06 only contain LF3
620 and LF4 shows that these core sites are located in a relatively high-energy
621 depositional environment, where such events are frequent, which is a logical
622 consequence of their relatively ice-proximal location.

623 Although we have no lithological evidence from Ymerbukta and Trygghamna,
624 we suggest that similar sedimentary processes to those in Magdaleneforden also
625 operate in the two Isfjorden tributaries. This is based on the sub-bottom profiler
626 data, showing that sedimentary facies in the three fjords are very similar, and
627 that sedimentation is controlled by glacial meltwater, occasional icebergs, and
628 occasional to frequent gravitational mass flows (Forwick & Vorren, 2011, 2012,
629 this study). This corroborates the findings of previous studies, which showed
630 that the sedimentary processes in front of tidewater glaciers are generally sim-
631 ilar, and that it is more the magnitude and the different factors controlling
632 sedimentation rather than the individual processes themselves that vary with

633 climate and locality (e.g. Elverhøi et al., 1983; Cowan et al., 1999; Ó Cofaigh
634 & Dowdeswell, 2001; Ó Cofaigh et al., 2001; Gilbert et al., 2002; Streuff et al.,
635 2017).

636 **5.3. Evolution of the sediment-landform assemblages and** 637 **associated glacier dynamics**

638 From the different landform assemblages observed in the three fjords, we infer
639 regionally variable glacier dynamics. In Ymerbukta, the Esmarkbreen Assem-
640 blage is consistent with submarine landform assemblages deposited from glacier
641 surges (e.g. Solheim & Pfirman, 1985; Ottesen et al., 2008; Flink et al., 2015;
642 Streuff et al., 2015), not least because the crevasse-squeeze component in the De
643 Geer moraines may be indicative of surge activity (cf. e.g. Sharp, 1985; Otte-
644 sen & Dowdeswell, 2006; Farnsworth et al., 2016). The presence of overridden
645 moraines indicates that Esmarkbreen experienced at least one re-advance after
646 deglaciation. Given that Esmarkbreen was not part of the original surge-type
647 glacier inventory put forward by Hagen et al. in 1993, but was recently identi-
648 fied as surge-type based on the presence of crevasse-squeeze ridges observed in
649 the glacier’s terrestrial foreland (Farnsworth et al., 2016), Esmarkbreen prob-
650 ably only surged recently (2014; Allaart et al., 2015), suggesting that most of
651 the landforms in the Ymerbukta assemblage are relatively young. The excep-
652 tions are the overridden moraines, that must have formed prior to the surge
653 and thus provide evidence for episodic ice retreat during deglaciation or af-
654 ter a LIA advance of Esmarkbreen. A later re-advance, likely related to a
655 surge of Esmarkbreen, formed the glacial lineations during the active phase of
656 rapid advance, the terminal moraine and the associated debris lobe during the
657 transition from active to passive phase, and a sequence of De Geer moraines
658 during subsequent episodic retreat. If we assume that the De Geer moraines

were formed each year during winter as suggested for other Spitsbergen fjords (e.g. Ottesen & Dowdeswell, 2006; Ottesen et al., 2008) the total number of such moraines in Ymerbukta should indicate that Esmarkbreen surged around 50 years ago. However, given the crevasse-squeeze component of some of these ridges, the fact that Esmarkbreen did not show surge-like behaviour until ~2010 (Heidi Sevestre, pers. comm.), and the comparatively small number of De Geer moraines in Trygghamna and Magdalenefjorden (see below), we suggest that the moraines form much less regularly than previously thought and may not necessarily be related to winter re-advances.

Glacier dynamics such as those suggested here for Esmarkbreen have been documented for a large number of Spitsbergen surge-type glaciers (cf. Solheim & Pfirman, 1985; Solheim, 1991; Boulton et al., 1996; Christoffersen et al., 2005; Ottesen et al., 2008; Kristensen et al., 2009; Flink et al., 2015; Streuff et al., 2015), supporting the hypothesis that Esmarkbreen is indeed a surging glacier (Farnsworth et al., 2016). The sub-bottom profiler data from the fjord shows that sedimentation was largely similar throughout the Holocene with mud settling from glacial meltwater, IRD melting out from icebergs and or sea ice, and occasional mass transport events reworking the accumulated deposits (Forwick & Vorren, 2009, 2011).

In Trygghamna we identified 29 De Geer moraines and a large debris lobe on the flank of the outermost moraine (see Figs. 3, 4d) as part of the Kjerulfbreen-Harrietbreen Assemblage. This outermost moraine could represent a terminal moraine marking the Holocene maximum ice extent in the fjord. Despite the similarities in landforms between Trygghamna and Ymerbukta, and although both Kjerulfbreen and Harrietbreen at the head of Trygghamna have recently been identified as surge-type glaciers (Farnsworth et al., 2016), we hesitate to ascribe a definitive surge origin to the submarine landforms in Trygghamna.

686 This is mainly because the absence of overridden moraines indicates only one
 687 Holocene re-advance of the glaciers in the fjord, which could also be related
 688 to the LIA cooling rather than a surge (cf. e.g. Plassen et al., 2004; Ottesen
 689 & Dowdeswell, 2006; Forwick et al., 2010), and because the absence of glacial
 690 lineations suggests relatively slow ice flow (cf. King et al., 2009), which is atypical
 691 for the active phase of a surging glacier (e.g. Meier & Post, 1969; Raymond, 1987;
 692 Sharp, 1988; Benn & Evans, 2010). The landform assemblage in Trygghamna
 693 could thus be a product of ice retreat, either during deglaciation or after the
 694 LIA, with the outermost moraine indicative of a prolonged period of still-stand
 695 related to either (i) shoaling waters and accordingly slowed glacier retreat (e.g.
 696 Oerlemans & Nick, 2006; Benn et al., 2007; Kehrl et al., 2011), or (ii) the
 697 transition between ice advance and ice retreat during the LIA. Nevertheless,
 698 if crevasse-squeeze ridges are indeed surge-diagnostic landforms, their presence
 699 in the terrestrial forelands of both glaciers (Farnsworth et al., 2016), and the
 700 variable orientations of some of the submarine De Geer moraines should suggest
 701 that the entire landform assemblage formed from a glacier surge. Based on the
 702 absence of overridden moraines and glacial lineations, we suggest that formation
 703 of the terminal moraine and the associated debris lobe in Trygghamna occurred
 704 as a result of a re-advance of Harrietbreen and Kjerulfbreen during the LIA
 705 cooling (cf. Ottesen & Dowdeswell, 2006), but note that glacier advance and
 706 subsequent landform formation could have occurred from both glacier surging
 707 or LIA advances. Similar conclusions have also been drawn for the adjacent
 708 Borebukta and Yoldiabukta (Ottesen & Dowdeswell, 2006).

709 The debris lobes and pockmarks in the outer parts of both Ymerbukta and
 710 Trygghamna must have formed during the Holocene after the fjord was ice-
 711 free, as a grounded glacier would have prevented the deposition of large mass-
 712 transport deposits and the escaping of fluids from the seafloor (cf. Forwick et al.,

2009; Forwick & Vorren, 2012).

Glacial lineations in Isfjorden are transverse to the direction of inferred ice flow through Ymerbukta and Trygghamna and are thus likely the product of ice streaming east to west through the fjord during the LGM (cf. e.g. Landvik et al., 1998; Ottesen et al., 2005). The ice-marginal wedge provides evidence for an extended period of still-stand during overall glacier retreat.

In Magdalenefjorden, the presence of a deep bathymetric basin in the central fjord complicates the reconstruction of the local ice dynamics, as it is difficult to ascertain whether a lack of glacial landforms within the central parts of this basin is related to (i) partial ungrounding of the retreating ice margin during deglaciation, or (ii) post-glacial sediment masking. The De Geer moraines in the fjord could thus have formed during episodic ice retreat related to either deglaciation or a LIA re-advance of the local glaciers. As the moraines or parts thereof occur within the entire fjord and along the steep side walls of the bathymetric basin, it is likely that at least the moraines in the outer fjord, and possibly those within the basin, were deposited during deglaciation when ice retreated from its maximum position at the continental shelf edge (cf. Ottesen & Dowdeswell, 2009). In the inner fjord, the moraines could also be a product of Holocene glacier re-advance and subsequent retreat. Indeed, the most distal of the De Geer moraines here is slightly larger than the rest and was previously interpreted as a terminal moraine marking the maximum ice extent of Waggonwaybreen during the LIA (Ottesen & Dowdeswell, 2009). This interpretation is supported by our lithological evidence and sub-bottom profiler data which show the presence of a small and localised debris flow down the moraine's distal flank. We note, that in this instance the assumption of annual formation for the De Geer moraines would also be unreasonable, as the total number of 37 would suggest that Waggonwaybreen reached its LIA maximum less than 40

740 years ago.

741

742 From the lithological data in Madalenefjorden we summarise the following
743 chronology for the sediments in the fjord: LD2a is the oldest recovered facies in
744 the cores and was deposited during relatively ice-distal conditions. Based also
745 on the AMS date of around 500 cal a BP, it is likely that this subfacies was
746 deposited during glacier retreat during deglaciation, probably before the LIA.
747 The fact that GC03 mostly contains LF2a shows that the sedimentary processes
748 at this core site were largely the same throughout the Holocene. In GC01 LF2a
749 is overlain by LF2b, which was interpreted as proximal glacimarine sediment.
750 The stratification, the high SAR and the accumulation of this facies relatively
751 soon after ~ 320 cal a BP could suggest that Waggonwaybreen re-advanced in
752 response to the LIA cooling. The increasingly more diffuse stratification up-
753 core of GC01 and the re-appearance of LF2a before ~ 230 cal a BP implies that
754 retreat after the LIA advance was already underway at this point. The ice-distal
755 sediments of LF1 at the top of both GC01 and GC03 indicate that the glacier
756 front had retreated relatively far back from the core sites by ~ 230 cal a BP.

757 Lithofacies LF3 reflects the contemporary sedimentary environment at the
758 core site of GC06, where the input from many different sources maintains a high
759 sediment accumulation rate and causes occasional small changes in the deposi-
760 tional environment. Since LF4 was interpreted as deposits from gravitational
761 mass-flow events, this lithofacies is chronologically independent, as such events
762 can occur at any time. Nevertheless, where the facies occurs as thin beds in the
763 mud of LF2b in GC01, the deposition of LF4 was likely related to the possible
764 re-advance of Waggonwaybreen during the LIA.

765 6. Conclusions

766 Swath-bathymetric data from three fjords reveal the landform assemblages in
767 front of several Spitsbergen tidewater glaciers. While the main processes form-
768 ing the landforms are somewhat similar, individual glacier dynamics control
769 the occurrence of different types of landforms. In Ymerbukta, one or more
770 surges of Esmarkbreen are indicated by the presence of overridden moraines
771 and glacial lineations, formed during retreat followed by relatively fast glacier
772 re-advance. A terminal moraine and associated debris lobe mark the maximum
773 extent of the glacier during the Holocene, and the occurrence of numerous De
774 Geer moraines provides evidence for subsequent step-wise retreat. Processes of
775 crevasse-squeezing are implied by the variable orientation and occasional cross-
776 cutting of these moraines. In the adjacent Trygghamna a terminal moraine
777 and debris lobe were probably formed as a result of a LIA re-advance of the
778 two glaciers Kjerulfbreen and Harrietbreen at the head of the fjord, after which
779 retreat was similarly episodic as in Ymerbukta.

780 In Magdalenefjorden in northeast Spitsbergen a slightly larger moraine at
781 the edge of the innermost basin is probably related to a re-advance of Wag-
782 gonwaybreen in response to the LIA around 300 cal a BP. The characteristic
783 debris lobe often associated with the distal flank of such terminal moraines in
784 Svalbard fjords appears to be unusually small in Magdalenefjorden, suggesting
785 that the glacier front halted there for only a short period of time, or that sed-
786 iment availability was restricted. Retreat, as indicated by numerous De Geer
787 moraines, was also episodic. Our data interpretation indicates, that contrary
788 to previous assumptions, De Geer moraines can form much less frequently than
789 once a year. The transition from distal to proximal to distal glacimarine mud in
790 one of the gravity cores from Magdalenefjorden reflects a sequence of glacier re-
791 treat, likely related to deglaciation after the LGM, glacier re-advance probably

792 related to the LIA cooling, and subsequent retreat of the glacier front. Distal
793 glacimarine muds are massive to weakly stratified with occasional evidence of
794 bioturbation, and were deposited at relatively low rates of 0.04–0.49 cm a⁻¹.
795 Ice-proximal glacimarine muds are distinctly laminated or stratified and contain
796 couplets of one coarser and one finer layer, which probably derive from season-
797 ally controlled suspension rainout from meltwater plumes. The occurrence of
798 turbidity currents is indicated by the presence of several thin sandy layers in
799 the glacimarine mud. The proximal stratified muds accumulated at a rate of
800 around 3 cm a⁻¹.

801 The lithofacies observed in the gravity cores reveal that suspension settling
802 from meltwater plumes and the water column is the dominant sedimentary pro-
803 cess in Magdalenefjorden, with very occasional input of IRD from icebergs and
804 sea ice. These sediments are reworked as a consequence of turbidity currents
805 close to the glacier fronts and of other mass-flow events such as slides and slumps
806 down steep submarine slopes. Our data interpretation suggests that components
807 of landform assemblages as well as sedimentary processes in glacimarine envi-
808 ronments are largely similar and that small variations across wider areas are
809 controlled by localised changes in glacier hydrology, thermal regime, sediment
810 availability, bathymetry and air/ocean temperatures, rather than different geo-
811 graphic location and climate.

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Allaart, L., Håkansson, L., & Noormets, R. (2015). Field guide GLANAM
3rd workshop and field excursion in Isfjorden, Svalbard. URL:
http://www.glanam.org/wp-content/uploads/2017/03/Fieldguide_3AnnualMeeting_Ymerbukta_2015.pdf.

Baeten, N., Forwick, M., Vogt, C., & Vorren, T. (2010). Late Weichselian
and Holocene sedimentary environments and glacial activity in Billefjorden,
Svalbard. *Geological Society, London, Special Publications*, 344, 207–23.
doi:10.1144/SP344.15.

Batchelor, C., & Dowdeswell, J. (2015). Ice-sheet grounding-zone wedges
(GZWs) on high-latitude continental margins. *Marine Geology*, 363, 65–92.
doi:10.1016/j.margeo.2015.02.001.

Benn, D., & Evans, D. (2010). *Glaciers and Glaciation*. (2nd ed.). Hodder
Education.

Benn, D., Warren, C., & Mottram, R. (2007). Calving processes and the dy-
namics of calving glaciers. *Earth-Science Reviews*, 82, 143–79. doi:10.1016/
j.earscirev.2007.02.002.

Błaszczczyk, M., Jania, J. A., & Hagen, J. O. (2009). Tidewater glaciers of Sval-
bard: Recent changes and estimates of calving fluxes. *Polish Polar Research*,
30, 85–142.

Blatter, H., & Hutter, K. (1991). Polythermal conditions in Arctic glaciers.
Journal of Glaciology, 37, 261–9. doi:10.3198/1991JoG37-126-261-269.

- Boulton, G. (1986). Push-moraines and glacier-contact fans in marine and terrestrial environments. *Sedimentology*, 33, 667–98. doi:10.1111/j.1365-3091.1986.tb01969.x.
- Boulton, G., Van Der Meer, J., Hart, J., Beets, D., Ruegg, G., Van Der Wateren, F., & Jarvis, J. (1996). Till and moraine emplacement in a deforming bed surge – an example from a marine environment. *Quaternary Science Reviews*, 15, 961–87. doi:10.1016/0277-3791(95)00091-7.
- Christoffersen, P., Piotrowski, J. A., & Larsen, N. K. (2005). Basal processes beneath an Arctic glacier and their geomorphic imprint after a surge, Elisebreen, Svalbard. *Quaternary Research*, 64, 125–37. doi:10.1016/j.yqres.2005.05.009.
- Cowan, E., Cai, J., Powell, R., Clark, J., & Pitcher, J. (1997). Temperate glacial-marine varves: an example from Disenchantment Bay, southern Alaska. *Journal of Sedimentary Research*, 67, 536–49.
- Cowan, E., & Powell, R. (1990). Suspended sediment transport and deposition of cyclically interlaminated sediment in a temperate glacial fjord, Alaska, USA. *Geological Society, London, Special Publications*, 53, 75–89. doi:10.1144/GSL.SP.1990.053.01.04.
- Cowan, E. A., Seramur, K. C., Cai, J., & Powell, R. D. (1999). Cyclic sedimentation produced by fluctuations in meltwater discharge, tides and marine productivity in an Alaskan fjord. *Sedimentology*, 46, 1109–26. doi:10.1046/j.1365-3091.1999.00267.x.
- Dallmann, W., Ohta, Y., Elvevold, S., & Blomeier, D. (2002). *Bedrock map of Svalbard and Jan Mayen*. Norsk Polarinstitut, Temakart No. 33.

- Dowdeswell, J. A., Elverhøi, A., & Spielhagen, R. (1998). Glacimarine sedimentary processes and facies on the Polar North Atlantic margins. *Quaternary Science Reviews*, 17, 243–72. doi:10.1016/S0277-3791(97)00071-1.
- Dowdeswell, J. A., & Ottesen, D. (2016). Eskers formed at the beds of modern surge-type tidewater glaciers in Spitsbergen. In J. A. Dowdeswell, M. Canals, M. Jakobsson, B. J. Todd, E. K. Dowdeswell, & K. A. Hogan (Eds.), *Atlas of submarine glacial landforms* number 46 in Geological Society Memoir (pp. 83–4). The Geological Society London. doi:10.1144/M46.70.
- Dowdeswell, J. A., Whittington, R., Jennings, A., Andrews, J., Mackensen, A., & Marienfeld, P. (2000). An origin for laminated glacimarine sediments through sea-ice build-up and suppressed iceberg rafting. *Sedimentology*, 47. doi:10.1046/j.1365-3091.2000.00306.x.
- Elverhøi, A., Andersen, E., Dokken, T., Hebbeln, D., Spielhagen, R., Svendsen, J., & Forsberg, C. (1995). The growth and decay of the Late Weichselian ice sheet in western Svalbard and adjacent areas based on provenance studies of marine sediments. *Quaternary Research*, 44, 303–16. doi:10.1006/qres.1995.1076.
- Elverhøi, A., Liestøl, O., & Nagy, J. (1980). Glacial erosion, sedimentation and microfauna in the inner part of Kongsfjorden, Spitsbergen. *Norsk Polarinstitutt Skrifter*, 172, 33–58.
- Elverhøi, A., Lønne, Ø., & Seland, R. (1983). Glaciomarine sedimentation in a modern fjord environment, Spitsbergen. *Polar Research*, 1, 127–50. doi:10.1111/j.1751-8369.1983.tb00697.x.
- Evans, D., & Orton, C. (2014). Heinabergsjökull and Skalafellsjökull, Iceland: active temperate piedmont lobe and outwash head glacial landsystem. *Journal of Maps*, 11, 415–31. doi:10.1080/17445647.2014.919617.

- Farnsworth, W. R., Ingólfsson, Ó., Retelle, M., & Schomacker, A. (2016). Over 400 previously undocumented Svalbard surge-type glaciers identified. *Geomorphology*, *264*, 52–60. doi:10.1144/GSL.SP.2002.203.01.09.
- Flink, A. E., Noormets, R., Fransner, O., Hogan, K. A., O'Regan, M., & Jakobsson, M. (2017). Past ice flow in Wahlenbergfjorden and its implications for late Quaternary ice sheet dynamics in northeastern Svalbard. *Quaternary Science Reviews*, .
- Flink, A. E., Noormets, R., Kirchner, N., Benn, D. I., Luckman, A., & Lovell, H. (2015). The evolution of a submarine landform record following recent and multiple surges of Tunabreen glacier, Svalbard. *Quaternary Science Reviews*, *108*, 37–50. doi:10.1016/j.quascirev.2014.11.006.
- Forman, S. (1990). Post-glacial relative sea-level history of northwestern Spitsbergen, Svalbard. *Geological Society of America Bulletin*, *102*, 1580–90. doi:10.1130/0016-7606(1990)102<1580:PGRSLH>2.3.CO;2.
- Forwick, M., Baeten, N., & Vorren, T. (2009). Pockmarks in Spitsbergen fjords. *Norwegian Journal of Geology*, *89*, 65–77.
- Forwick, M., & Vorren, T. (2007). Holocene mass-transport activity and climate in outer Isfjorden, Spitsbergen: marine and subsurface evidence. *The Holocene*, *17*(6), 707–16. doi:10.1177/0959683607080510.
- Forwick, M., & Vorren, T. (2009). Late Weichselian and Holocene sedimentary environments and ice rafting in Isfjorden, Spitsbergen. *Paleogeography, Paleoclimatology, Paleoecology*, *280*(1), 258–74. doi:10.1016/j.palaeo.2009.06.026.
- Forwick, M., & Vorren, T. (2011). Stratigraphy and deglaciation of the Isfjorden area, Spitsbergen. *Norwegian Journal of Geology*, *90*, 163–79.

- Forwick, M., Vorren, T., Hald, M., Korsun, S., Roh, Y., Vogt, C., & Yoo, K. (2010). Spatial and temporal influence of glaciers and rivers on the sedimentary environment in Sassenfjorden and Tempelfjorden, Spitsbergen. *Geological Society, London, Special Publications*, *344*(1), 163–93. doi:10.1144/SP344.13.
- Forwick, M., & Vorren, T. O. (2012). Submarine Mass Wasting in Isfjorden, Spitsbergen. In Y. Yamada, K. Kawamura, K. Ikehara, Y. Ogawa, R. Urgeles, D. Mosher, J. Chaytor, & M. Strasser (Eds.), *Submarine mass movements and their consequences* (pp. 711–22). Springer Springer+ Business Media volume 31. doi:10.1007/978-94-007-2162-3\underline{63}.
- Fransner, O., Noormets, R., Flink, A. E., Hogan, K., O'Regan, M., & Jakobsson, M. (2017). Glacial landforms and their implications for glacier dynamics in Rijpfjorden and Duvefjorden, northern Nordaustlandet, Svalbard. *Journal of Quaternary Science*, .
- Gilbert, R. (1982). Contemporary sedimentary environments on Baffin Island, NWT, Canada: Glaciomarine processes in fiords of eastern Cumberland Peninsula. *Arctic and Alpine Research*, *14*, 1–12. doi:10.2307/1550809.
- Gilbert, R., Aitken, A., & Lemmen, D. (1993). The glaciomarine sedimentary environment of Expedition Fiord, Canadian High Arctic. *Marine Geology*, *110*, 257–73. doi:10.1016/0025-3227(93)90088-D.
- Gilbert, R., Nielsen, N., Möller, H., Desloges, J., & Rasch, M. (2002). Glaciomarine sedimentation in Kangerdluk (Disko Fjord), West Greenland, in response to a surging glacier. *Marine Geology*, *191*, 1–18. doi:10.1016/S0025-3227(02)00543-1.
- Hagen, J. (1993). *Glacier atlas of Svalbard and Jan Mayen* volume 129. Norsk Polarinstitut Middelelser. URL: <http://hdl.handle.net/11250/173065>.

- Hagen, J. O., Kohler, J., Melvold, K., & Winther, J.-G. (2003). Glaciers in Svalbard: mass balance, runoff and freshwater flux. *Polar Research*, 22, 145–59. doi:10.1111/j.1751-8369.2003.tb00104.x.
- Hogan, K. A., Dowdeswell, J. A., Noormets, R., Evans, J., & Ó Cofaigh, C. (2010). Evidence for full-glacial flow and retreat of the Late Weichselian Ice Sheet from the waters around Kong Karls Land, eastern Svalbard. *Quaternary Science Reviews*, 29, 3563–82. doi:10.1016/j.quascirev.2010.05.026.
- Holland, D., Thomas, R., De Young, B., Ribergaard, M., & Lyberth, B. (2008). Acceleration of Jakobshavn Isbræ triggered by warm subsurface ocean waters. *Nature Geoscience*, 1, 659–64. doi:10.1038/ngeo316.
- Hovland, M., & Judd, A. G. (1988). *Seabed pockmarks and seepages: impact on geology, biology and the marine environment*.
- Ingólfsson, O., & Landvik, J. (2013). The Svalbard-Barents Sea ice-sheet – historical, current and future perspectives. *Quaternary Science Reviews*, 64, 33–60. doi:10.1016/j.quascirev.2012.11.034.
- Jennings, A. E., & Weiner, N. J. (1996). Environmental change in eastern Greenland during the last 1300 years: evidence from foraminifera and lithofacies in Nansen fjord, 68 °N. *The Holocene*, 6, 179–91. doi:10.1177/095968369600600205.
- Jessen, S., Rasmussen, T., Nielsen, T., & Solheim, A. (2010). A new Late Weichselian and Holocene marine chronology for the western Svalbard slope 30000-0 cal years BP. *Quaternary Science Reviews*, 29(9), 1301–12. doi:10.1016/j.quascirev.2010.02.020.
- Kehrl, L., Hawley, R., Powell, R., & Brigham-Grette, J. (2011). Glacimarine

- sedimentation processes at Kronebreen and Kongsvegen, Svalbard. *Journal of Glaciology*, 57(205), 841–7. doi:10.3189/002214311798043708.
- Kempf, P., Forwick, M., Laberg, J., & Vorren, T. (2013). Late Weichselian – Holocene sedimentary palaeoenvironment and glacial activity in the high-Arctic van Keulenfjorden, Spitsbergen. *The Holocene*, 23(11), 1607–18. doi:10.1177/0959683613499055.
- King, E., Hindmarsh, R., & Stokes, C. (2009). Formation of mega-scale glacial lineations observed beneath a West Antarctic ice stream. *Nature Geoscience*, 2(8), 585–8. doi:10.1038/ngeo581.
- Koç, N., Klitgaard-Kristensen, D., Hasle, K., Forsberg, C., & Solheim, A. (2002). Late glacial palaeoceanography of Hinlopen Strait, northern Svalbard. *Polar Research*, 21, 307–14. doi:10.1111/j.1751-8369.2002.tb00085.x.
- Kohler, J., James, T. D., Murray, C. B. O., T. and Nuth, Barrand, N., Aas, H., & Luckman, A. (2007). Acceleration in thinning rate on western Svalbard glaciers. *Geophysical Research Letters*, 34. doi:10.1029/2007GL030681.
- Kristensen, L., Benn, D., Holmes, A., & Ottesen, D. (2009). Mud aprons in front of Svalbard surge moraines: Evidence of subglacial deforming layers or proglacial glaciotectonics? *Geomorphology*, 111(3), 206–21. doi:10.1016/j.geomorph.2009.04.022.
- Kuenen, P. (1948). Slumping in the Carboniferous rocks of Pembrokeshire. *Quarterly Journal of the Geological Society*, 104, 365–85. doi:10.1144/GSL.JGS.1948.104.01-04.18.
- Landvik, J., Bondevik, S., Elverhøi, A., Fjeldskaar, W., Mangerud, J., Salvigsen, O., Siegert, M. J., Svendsen, J.-I., & O., V. T. (1998). The last glacial maximum of Svalbard and the Barents Sea area: ice sheet extent and configuration.

- Quaternary Science Reviews*, 17(1-3), 43–75. doi:10.1016/S0277-3791(97)00066-8.
- Landvik, J., Brook, E., Gualtieri, L., Raisbeck, G., Salvigsen, O., & Yiou, F. (2003). Northwest Svalbard during the last glaciation: Ice-free areas existed. *Geology*, 31, 905–8. doi:10.1130/G19703.1.
- Landvik, J., Ingólfsson, O., Mienert, J., Lehman, S., Solheim, A., Elverhøi, A., & Ottesen, D. (2005). Rethinking Late Weichselian ice-sheet dynamics in coastal NW-Svalbard. *Boreas*, 34(1), 7–24. doi:10.1111/j.1502-3885.2005.tb01001.x.
- Landvik, J. Y., Hjort, C., Mangerud, J., Moller, P., & Salvigsen, O. (1995). The Quaternary record of eastern Svalbard-an overview. *Polar Research*, 14, 95–104. doi:10.1111/j.1751-8369.1995.tb00683.x.
- Lønne, I., & Mangerud, J. (1991). An Early or Middle Weichselian sequence of proglacial, shallow marine sediments on western Svalbard. *Boreas*, 20, 85–104. doi:10.1111/j.1502-3885.1991.tb00298.x.
- Lovell, T. V. (2000). As climate changes, so do glaciers. *Proceedings of the National Academy of Sciences of the United States of America*, 97, 1351–4. doi:10.1073/pnas.97.4.1351.
- Lundqvist, J. (1981). Moraine morphology. Terminological remarks and regional aspects. *Geografiska Annaler. Series A. Physical Geography*, (pp. 127–38). doi:10.2307/520824.
- Mackiewicz, N., Powell, R., Carlson, P., & Molnia, B. (1984). Interlaminated ice-proximal glacimarine sediments in Muir Inlet, Alaska. *Marine Geology*, 57, 113–47. doi:10.1016/0025-3227(84)90197-X.

- Mangerud, J. (1972). Radiocarbon dating of marine shells, including a discussion of apparent age of recent shells from Norway. *Boreas*, 1, 143–72. doi:10.1111/j.1502-3885.1972.tb00147.x.
- Mangerud, J., Bolstad, M., Elgersma, A., Helliksen, D., Landvik, J. Y., Lønne, I., Lycke, A. K., Salvigsen, O., Sandahl, T., & Svendsen, J. I. (1992). The last glacial maximum on Spitsbergen, Svalbard. *Quaternary Research*, 38, 1–31. doi:10.1016/0033-5894(92)90027-G.
- Mangerud, J., & Landvik, J. (2007). Younger Dryas cirque glaciers in western Spitsbergen: smaller than during the Little Ice Age. *Boreas*, 36(3), 278–85. doi:10.1111/j.1502-3885.2007.tb01250.x.
- Meier, M., & Post, A. (1969). What are glacier surges? *Canadian Journal of Earth Sciences*, 6, 807–17. doi:10.1139/e69-081.
- Ó Cofaigh, C., & Dowdeswell, J. (2001). Laminated sediments in glacimarine environments: diagnostic criteria for their interpretation. *Quaternary Science Reviews*, 20, 1411–36. doi:10.1016/S0277-3791(00)00177-3.
- Ó Cofaigh, C., Dowdeswell, J., Allen, C., Hiemstra, J., Pudsey, C., Evans, J., & Evans, D. (2005). Flow dynamics and till genesis associated with a marine-based Antarctic paleo-ice stream. *Quaternary Science Reviews*, 24, 709–40. doi:10.1016/j.quascirev.2004.10.006.
- Ó Cofaigh, C., Dowdeswell, J., & Grobe, H. (2001). Holocene glacimarine sedimentation, inner Scoresby Sund, East Greenland: the influence of fast-flowing ice-sheet outlet glaciers. *Marine Geology*, 175, 103–29. doi:10.1016/S0025-3227(01)00117-7.
- Oerlemans, J., & Nick, F. (2006). Modelling the advance–retreat cycle of a tide-

- water glacier with simple sediment dynamics. *Global and Planetary Change*, 50, 148–60. doi:10.3189/172756407782871440.
- Ottesen, D., & Dowdeswell, J. (2006). Assemblages of submarine landforms produced by tidewater glaciers in Svalbard. *Journal of Geophysical Research*, 111, F01016. doi:10.1029/2005JF000330.
- Ottesen, D., & Dowdeswell, J. (2009). An inter-ice stream glaciated margin: submarine landforms and a geomorphic model based on marine-geophysical data from Svalbard. *Geological Society of America Bulletin*, 121(11/12), 1647–65. doi:10.1130/B26467.1.
- Ottesen, D., Dowdeswell, J., Benn, D., Kristensen, L., Christiansen, H., Christensen, O., & Vorren, T. (2008). Submarine landforms characteristic of glacier surges in two Spitsbergen fjords. *Quaternary Science Reviews*, 27(15), 1583–99. doi:10.1016/j.quascirev.2008.05.007.
- Ottesen, D., Dowdeswell, J., Landvik, J., & Mienert, J. (2007). Dynamics of the Late-Weichselian ice sheet on Svalbard inferred from high-resolution sea-floor morphology. *Boreas*, 36, 286–306. doi:10.1111/j.1502-3885.2007.tb01251.x.
- Ottesen, D., Dowdeswell, J., & Rise, L. (2005). Submarine landforms and the reconstruction of fast-flowing ice streams within a large Quaternary ice sheet: the 2500-km-long Norwegian-Svalbard margin (57°–80°N). *Geological Society of America Bulletin*, 117, 1033–50. doi:10.1130/B25577.1.
- Pettersson, R. (2004). *Dynamics of the cold surface layer of polythermal Stor-glaciären*. Ph.D. thesis Institutionen för naturgeografi och kvartärgeologi.
- Plassen, L., Vorren, T., & Forwick, M. (2004). Integrated acoustic and coring

- investigations of glacigenic deposits in Spitsbergen fjords. *Polar Research*, *23*(1), 89–110. doi:10.1111/j.1751-8369.2004.tb00132.x.
- Powell, R., & Molnia, B. (1989). Glacimarine sedimentary processes, facies and morphology of the south-southeast Alaska shelf and fjords. *Marine Geology*, *85*, 359–90. doi:10.1016/0025-3227(89)90160-6.
- Raymond, C. (1987). How do glaciers surge? A review. *Journal of Geophysical Research: Solid Earth (1978-2012)*, *92*(B9), 9121–34. doi:10.1029/JB092iB09p09121.
- Reimer, P., Bard, E., Bayliss, A., Beck, J., Blackwell, P., Ramsey, C., Buck, C., Cheng, H., Edwards, R., Friedrich, M., Grootes, P., Guilderson, T., Haffidason, H., Hajdas, I., Hatté, C., Heaton, T., Hoffmann, D., Hogg, A., Hughen, K., Kaiser, F., Kromer, B., Manning, S., Niu, M., Reimer, R., Richards, D., Scott, E., Southon, J., Staff, R., Turney, C., & van der Plicht, J. (2013). IntCal13 and Marine13 Radiocarbon Age Calibration Curves 0–50,000 Years cal BP. *Radiocarbon*, *55*, 1869–87. doi:10.2458/azu\$\text{underscore}\$js\$\text{underscore}\$rc.55.16947.
- Roy, S., Hovland, M., Noormets, R., & Olaussen, S. (2015). Seepage in Isfjorden and its tributary fjords, West Spitsbergen. *Marine Geology*, *363*, 146–59. doi:10.1144/GSL.SP.1993.070.01.06.
- Roy, S., Senger, K., Braathen, A., Noormets, R., Hovland, M., & Olaussen, S. (2014). Fluid migration pathways to seafloor seepage in inner Isfjorden and Adventfjorden, Svalbard. *Norwegian Journal of Geology*, *94*, 99–119. URL: <http://hdl.handle.net/1956/10661>.
- Senger, K., Roy, S., Braathen, A., Buckley, S. J., Bælum, K., Gernigon, L., Mjelde, R., Noormets, R., Ogata, K., Olaussen, S., Planke, S., Ruud, B., &

- Tveranger, J. (2013). Geometries of doleritic intrusions in central Spitsbergen, Svalbard: an integrated study of an onshore-offshore magmatic province with implications for CO₂ sequestration, . *93*, 143–6.
- Sexton, D. J., Dowdeswell, J. A., Solheim, A., & Elverhøi, A. (1992). Seismic architecture and sedimentation in northwest Spitsbergen fjords. *Marine Geology*, *103*, 53–68. doi:10.1016/0025-3227(92)90008-6.
- Sharp, M. (1985). Crevasse-fill-ridges - a landform type characteristic of surging glaciers? *Geografiska Annaler. Series A. Physical Geography*, *67*(3/4), 213–20. doi:10.2307/521099.
- Sharp, M. (1988). Surging glaciers: behaviour and mechanisms. *Progress in Physical Geography*, *12*(3), 349–70. doi:10.1177/030913338801200302.
- Solheim, A. (1991). The depositional environment of surging sub-polar tidewater glaciers. *Norsk Polarinstitutt Skrifter*, *194*, 1–97.
- Solheim, A., & Pfirman, S. (1985). Sea-floor morphology outside a grounded, surging glacier; Bråsvellbreen, Svalbard. *Marine Geology*, *65*, 127–43. doi:10.1016/0025-3227(85)90050-7.
- Stewart, F. S., & Stoker, M. S. (1990). Problems Associated with Seismic Facies Analysis of Diamicton-Dominated, Shelf Glacigenic Sequences. *Geo-Marine Letters*, *10*, 151–6. doi:10.1007/BF02085930.
- Streuff, K. (2013). *Landform assemblages in inner Kongsfjorden, Svalbard: Evidence of recent glacial (surge) activity*. Master's thesis University of Tromsø.
- Streuff, K., Forwick, M., Szczuciński, W., Andreassen, K., & Ó Cofaigh, C. (2015). Landform assemblages in inner Kongsfjorden, Svalbard: evidence of recent glacial (surge) activity. *Arktos*, *1*. doi:10.1007/s41063-015-0003-y.

- Streuff, K., Ó Cofaigh, C., Noormets, R., & Lloyd, J. (2017). Submarine landforms and glacimarine sedimentary processes in Lomfjorden, east Spitsbergen. *Marine Geology*, . doi:10.1016/j.margeo.2017.04.014.
- Svendsen, J., Elverhøi, A., & Mangerud, J. (1996). The retreat of the Barents Sea Ice Sheet on the western Svalbard margin. *Boreas*, 25, 244–56. doi:10.1111/j.1502-3885.1996.tb00640.x.
- Svendsen, J., Mangerud, J., Elverhøi, A., Solheim, A., & Schüttenhelm, R. (1992). The Late Weichselian glacial maximum on western Spitsbergen inferred from offshore sediment cores. *Marine Geology*, 104, 1–17. doi:10.1016/0025-3227(92)90081-R.
- Syvitski, J., Andrews, J., & Dowdeswell, J. (1996). Sediment deposition in an iceberg-dominated glacimarine environment, East Greenland: basin fill implications. *Global and Planetary Change*, 12, 251–70. doi:10.1016/0921-8181(95)00023-2.
- Syvitski, J. P. M. (1989). On the deposition of sediment within glacier-influenced fjords: oceanographic controls. *Marine Geology*, 85, 301–29. doi:10.1016/0025-3227(89)90158-8.
- Tulaczyk, S., Scherer, R., & Clark, C. (2001). A ploughing model for the origin of weak tills beneath ice streams: a qualitative treatment. *Quaternary International*, 86, 59–70. doi:10.1016/S1040-6182(01)00050-7.
- Velle, J. (2012). *Holocene sedimentary environments in Smeerenburgfjorden, Spitsbergen*. Master's thesis University of Tromsø.

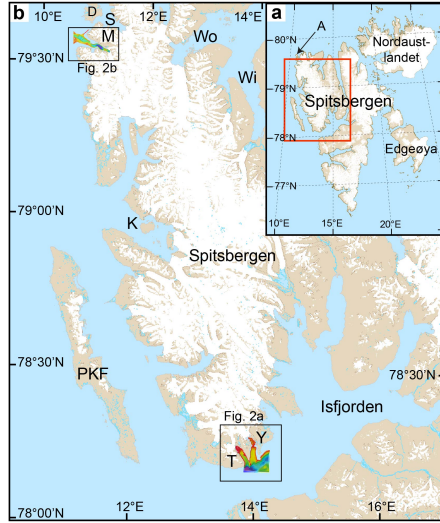


Figure 1: a) Overview of the Svalbard archipelago and its main islands. A = Amsterdamøya. The red rectangle shows the extent of b) map of the central and northwestern coast of Spitsbergen, with bathymetry coverage available for this study. Y = Ymerbukta, T = Trygghamna, PKF = Prins Karls Forland, K = Kongsfjorden, M = Magdalenefjorden, D = Danskøya, Wo = Woodfjorden, Wi = Wijdefjorden. Map data courtesy of the Norwegian Polar Institute.

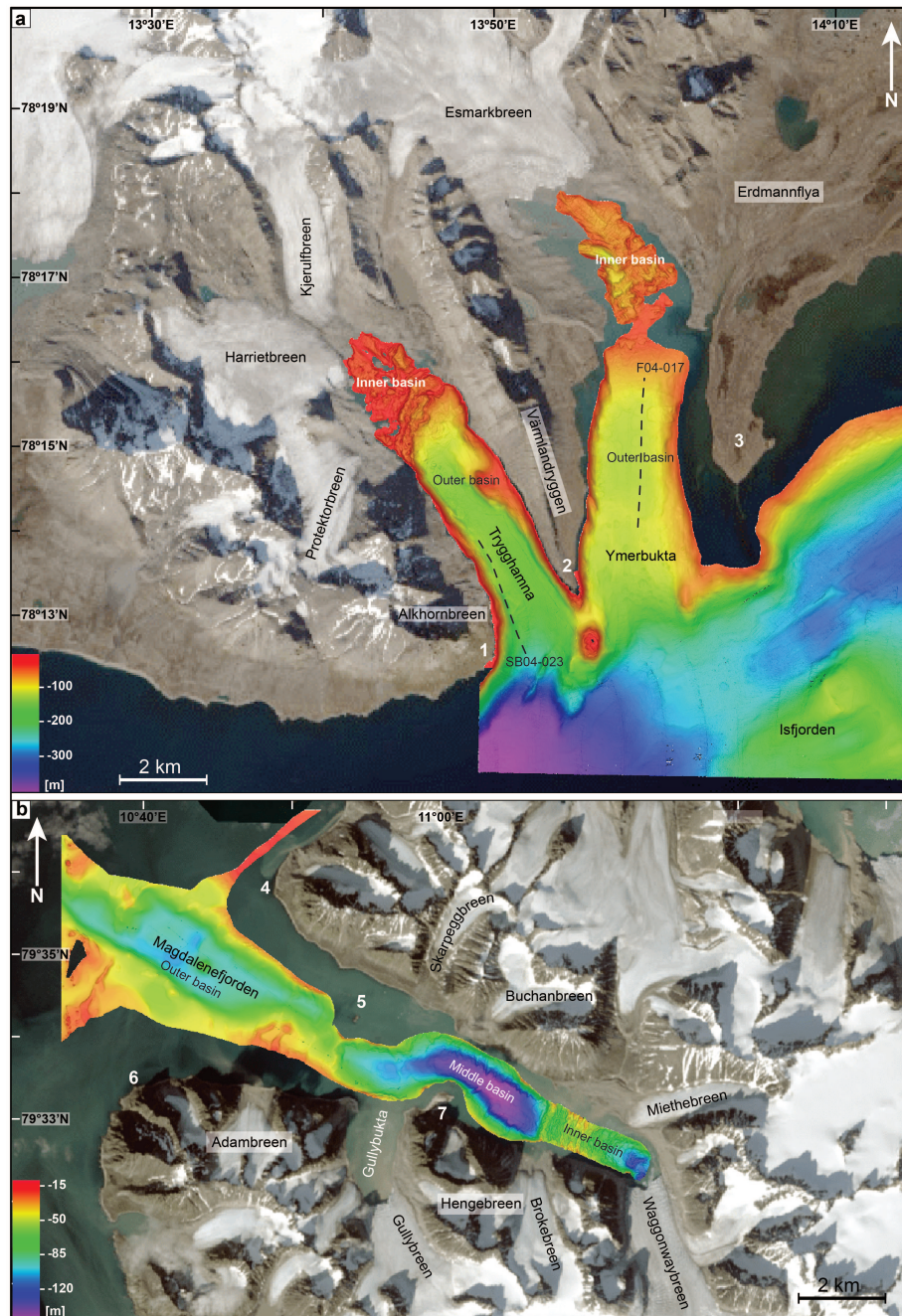


Figure 2: a) Ymerbukta and Trygghamna with their surroundings and the bathymetric data available from the two fjords. 1 = Alkepynten, 2 = Selmaneset, 3 = Erdmannodden. Dashed lines indicate the location of sparker and boomer profiles F04-163 and SB04-023, which were visualised and interpreted in Forwick & Vorren (2011, 2012) and provide the basis for the seismostratigraphy in Ymerbukta and Trygghamna. b) Magdalenefjorden, its surroundings, and the bathymetric data available for this study. 4 = Knattodden, 5 = Fugleholmen, 6 = Magdalenehuken, 7 = Gravneset. For an overview of the study areas' location see also Figure 1. The water depth scale is the same for all subsequent figures.

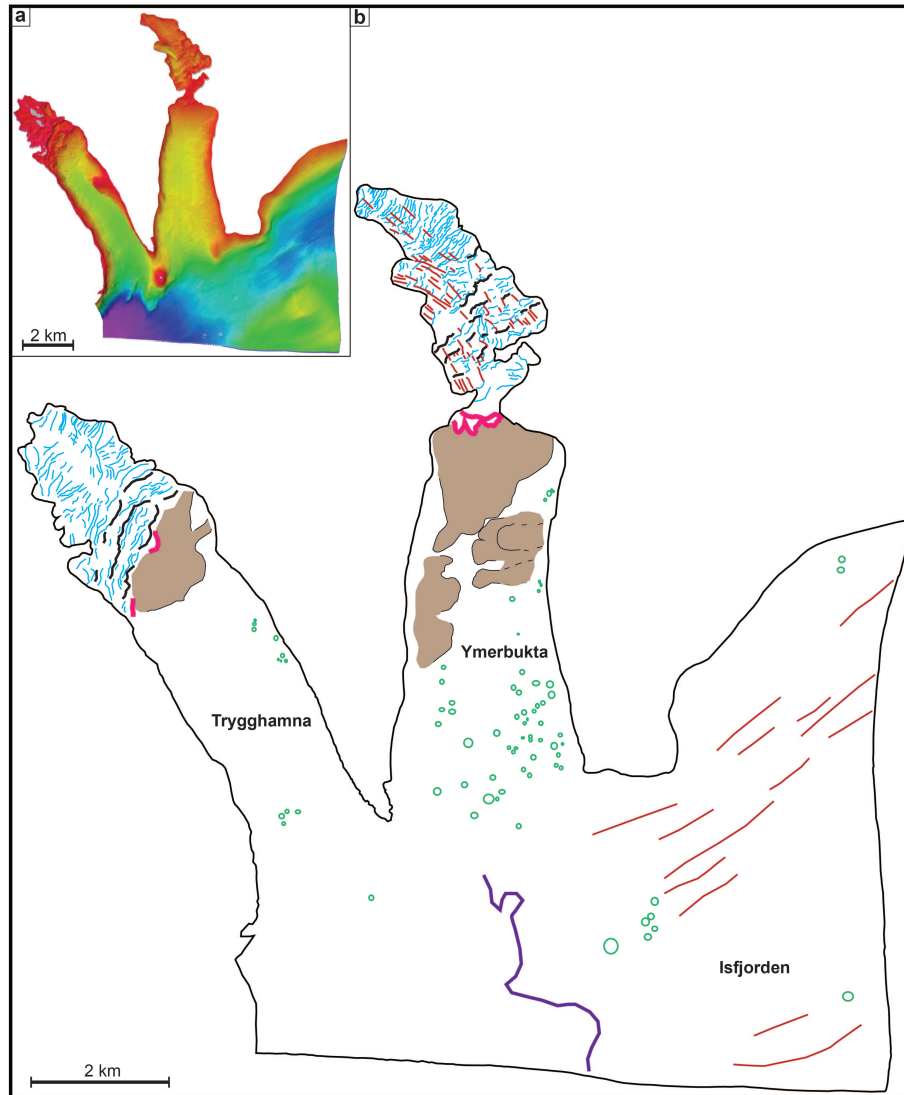


Figure 3: a) Swath-bathymetric data from Ymerbukta and Trygghamna. For the depth scale refer to Figure 2. b) Geomorphological map of the landforms in Ymerbukta, Trygghamna and parts of Isfjorden. Pockmarks and debris lobes were previously described and interpreted by Forwick et al. (2009); Forwick & Vorren (2012).

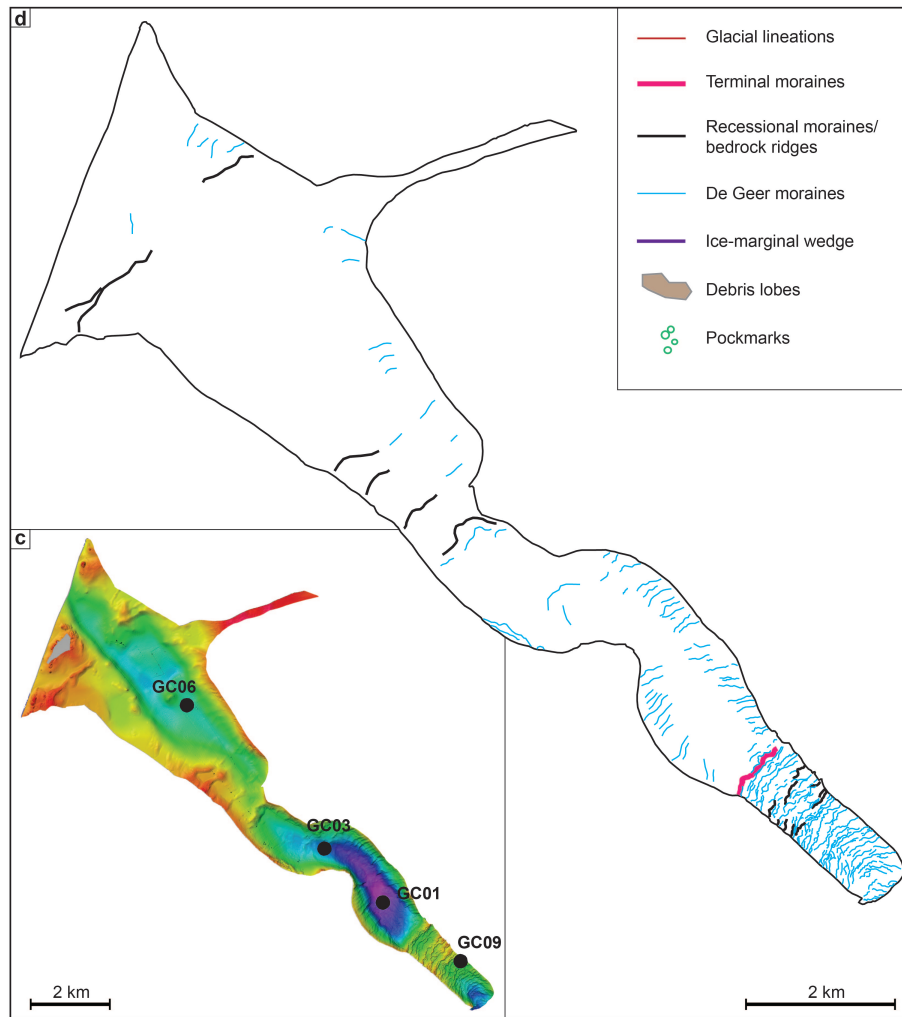


Figure 3 (cont.): c) Swath-bathymetric data with indicated core locations from Magdalenefjorden. For the water depth scale refer to Figure 2. d) Geomorphological map of the landforms in Magdalenefjorden. Part of the moraines/bedrock ridges in the outer and central fjord have been previously documented and interpreted by Ottesen & Dowdeswell (2009).

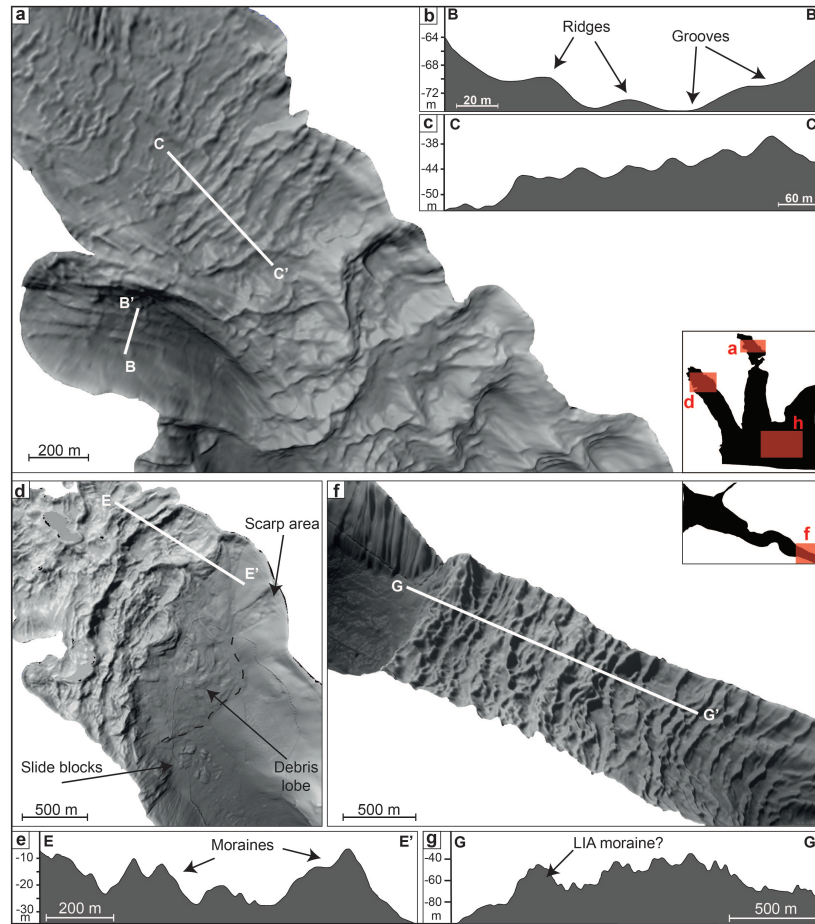


Figure 4: a) Seafloor of the inner fjord basin of Ymerbukta, showing examples of glacial lineations and De Geer moraines. Small red rectangles on the black polygon indicate the location of the subfigures a), d), and h). b) and c) show the cross-sectional profiles B–B' and C–C' across the landforms, respectively. d) Large transverse moraines and an associated debris lobe in Trygghamna. e) Cross-sectional profiles E–E' across Trygghamna moraines. f) De Geer moraines in Magdalenefjorden with g) cross-sectional profile G–G' across them.

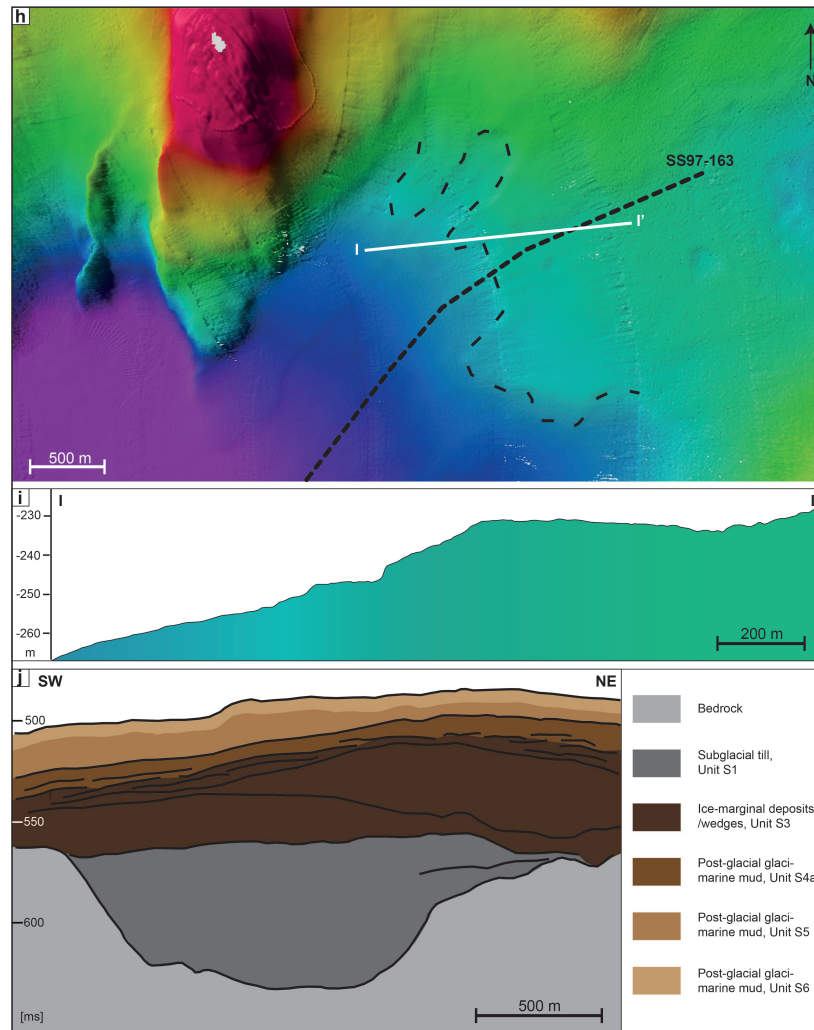


Figure 4 (cont.): h) Ice-marginal deposit in Isfjorden, with i) showing the cross-sectional profile I-I'. For water depth scale please refer to Figure 2. j) Acoustic facies interpretation of seismic line SS97-163 (based on, and modified after Forwick & Vorren, 2011). Y-axis shows two-way travel time in ms.

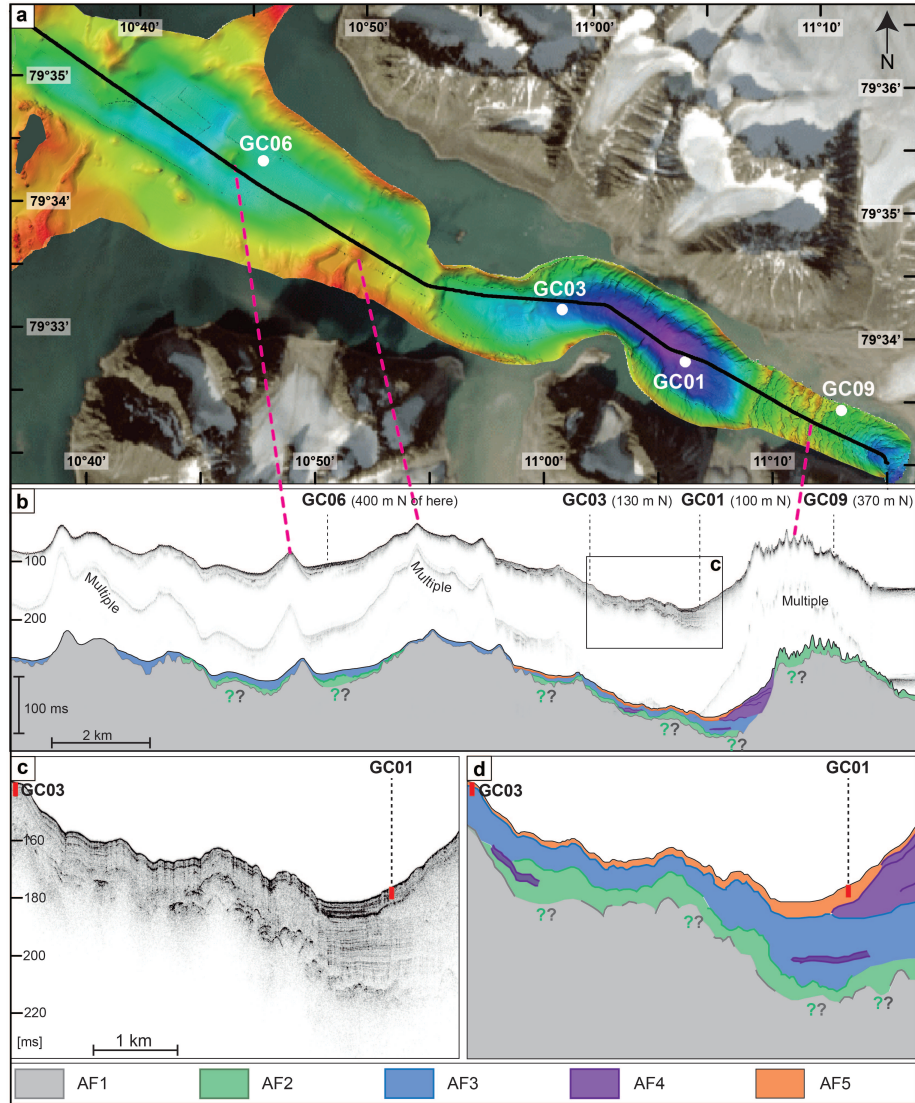


Figure 5: a) Bathymetry in Magdalenefjorden with the location of all cores (white dots) and chirp line 09JM-AG211-006 (black line) indicated. b) Seismic profile of chirp line 006, with approximate locations of the gravity cores. Pink stippled lines correlate bathymetric ridges with the according highs on the seismic profile. The interpretation of the acoustic facies in Magdalenefjorden is shown underneath the seismic profile. The black rectangle marks the extent of c). c) Detailed excerpt from chirp line 006 (black rectangle in b) showing examples of the acoustic facies in Magdalenefjorden. d) Acoustic facies interpretation of c). Red lines in c) and d) illustrate how far GC01 and GC03 penetrate into the seafloor sediments and how they correlate with the acoustic facies in the fjord. Y-axes show two-way travel time in ms.

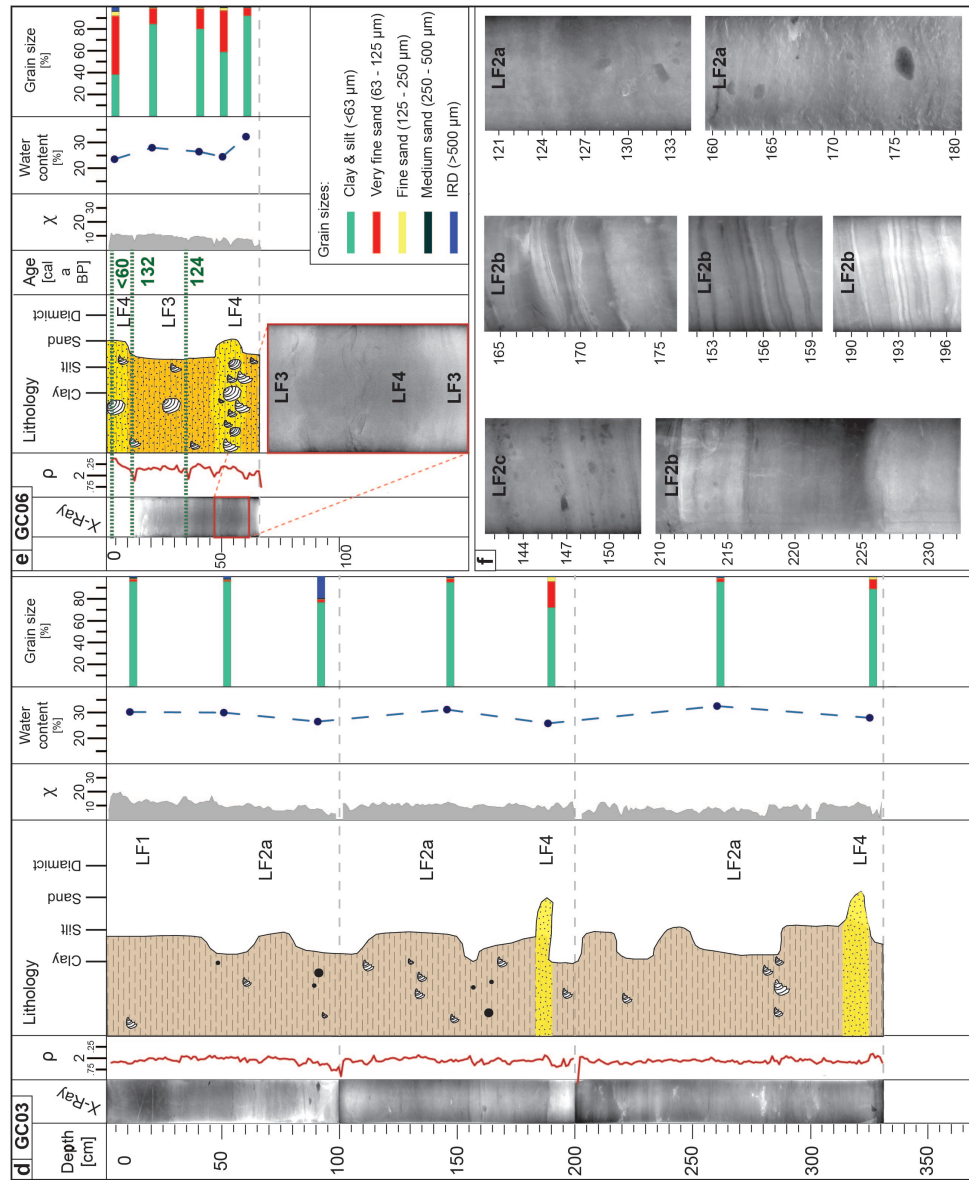


Figure 6 (cont.): d) X-radiographs, lithofacies log and physical properties of core GC03 in the central fjord. e) X-radiographs, lithofacies log and physical properties of core GC06 in the outer fjord. f) Examples of the x-radiographs showing the different types of laminated/stratified sediment in Magdalenefjorden.

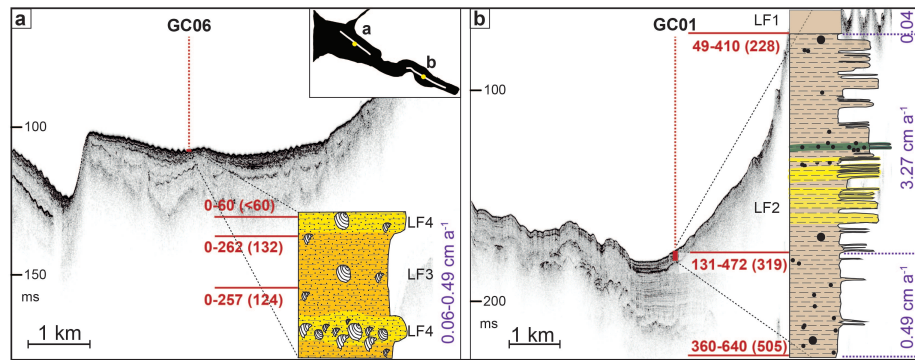


Figure 7: a) Detail of chirp line 09JM-AG211-003 with core location and log of GC06. White lines and dots on the black polygon in the top-right hand corner show the locations of a and b with respect to the bathymetric outline. b) Detail of chirp line 09JM-AG211-006 with the core location and log of GC01. Red lines and numbers show the 2- σ range and the median radiocarbon age in brackets at respective sediment depths, while purple labels refer to sediment accumulation rates. Y-axes show two-way travel time in ms.